HANDBOOK OF CLIMATOLOGY

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BY

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PART I. GENERAL CLIMATOLOGY

*TRANSLATED WITH THE AUTHOR'S PERMISSION
FROM THE SECOND REVISED AND ENLARGED GERMAN EDITION, WITH
ADDITIONAL REFERENCES AND NOTES, BY*

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PREFACE

THIS translation was undertaken primarily in order that it might serve as a text-book in the course in General Climatology in Harvard University. At the same time, the publication of an English edition of the standard work on climate, will, it is hoped, lead to the extension and improvement of the teaching of scientific climatology in the United States, as well as in Great Britain and her colonies.

The first edition of Professor Hann's *Handbuch der Klimatologie* appeared in 1883, in the "Bibliothek Geographischer Handbücher," edited by Professor Friedrich Ratzel. The second edition, in three volumes, was published in 1897. The first volume of the second German edition is the only one included in the present translation. This concerns General Climatology, and is a finished piece of work by itself. The last two volumes, which deal with Special Climatology, it has been found impracticable to translate.

This translation, as it stands, essentially reproduces the original. Numerous references, especially such as will be most useful to English and American students, have been added, and changes have been made in the text in order to bring the discussion down to date. A natural temptation to expand the original has been yielded to in very few cases only. Practically all of the important publications which have been issued since the completion of the second German edition are referred to. Some new examples of different climatic phenomena have been added, chiefly from the United States. Most of the examples given, however, necessarily still relate to Europe, because the climatology of that continent has been studied more critically than that of any other region. A few cuts have been made where the discussion concerned matters of special interest to European students only. Most of the paragraph headings are new, and the arrangement by parts, sections and chapters is somewhat different from that in the original. These

changes have been made with a view to adapting the book better for use in the class-room. Every change that has been made has the full approval of Professor Hann, who has been consulted in regard to all of these matters.

Every reference, the original of which is accessible in the Harvard College library or in the library of the Harvard College Observatory, has been looked up, verified, and made as complete as possible. No apology is needed for the use of the centigrade and metric systems in such a book as this. For convenience, conversion tables, reprinted from the Smithsonian Meteorological Tables, are given in the Appendix. In the tables, the figures indicating maxima are printed in heavy type, and those indicating minima are followed by an asterisk.

I am indebted first of all to Professor Hann, for the generous permission given me to translate his *Handbuch*; for his courtesy in allowing me to make so many changes and additions, and for the interest which he has shown throughout the progress of the translation. Further, I am under great obligations to Mr. Henry S. Mackintosh, of Keene, N.H., who has read both manuscript and proof, and to Mr. H. Helm Clayton, of Blue Hill Observatory, Hyde Park, Mass., who has read the manuscript. Both of these gentlemen have been of great help to me. Professor Frank W. Very has, with the approval of Professor Hann, made some slight revision of the sections on solar climate, on atmospheric absorption and radiation, and on geological changes of climate, with a view to bringing these discussions down to date. Professor Very has also been good enough to help me with the proof-reading. To my colleagues, Professor William M. Davis and Professor Robert W. Willson, I am also under obligations for assistance in various parts of the work.

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ABBREVIATIONS.

Z.f.M.—*Zeitschrift der österreichischen Gesellschaft für Meteorologie* (1866-1885).

M.Z.—*Meteorologische Zeitschrift* (1886 on).

(When these letters are enclosed in brackets they refer to an abstract or to a review.)

S.W.A., *D.W.A.*—*Sitzungsberichte* and *Denkschriften der kaiserlichen Akademie der Wissenschaften* (*Wien*). *Mathematisch-naturwissenschaftliche Classe.*

PART I.

THE CLIMATIC FACTORS

INTRODUCTION.

CLIMATOLOGY: ITS MEANING, AIMS, AND METHODS.[1]

Climate as distinguished from weather.—By *climate* we mean the sum total of the meteorological phenomena that characterise the average condition of the atmosphere at any one place on the earth's surface. That which we call *weather* is only one phase in the succession of phenomena whose complete cycle, recurring with greater or less uniformity every year, constitutes the climate of any locality. *Climate* is the sum total of the *weather* as usually experienced during a longer or shorter period of time at any given season. An account of a climate, therefore, means a description of the average state of the atmosphere. To illustrate, we may say "the weather in central Europe was very cold in December, 1879," or, "the weather was rainy in August, 1880." It is also correct to say that the climate of England is mild and damp in December, although December of the year 1879 happened to be very cold there. It is, however, a wrong use of the word climate when we say "the *climate* of Germany was rainy in the summer of 1882," for as soon as we speak of the atmospheric conditions of a single period of time, we must use the word *weather*.

Aims of climatology.—It is the object of climatology to make us familiar with the average conditions of the atmosphere in different parts of the earth's surface, as well as to inform us concerning any departures from these conditions which may occur at the same place during long intervals of time. Brevity demands that in the description of the climate of any place, only those weather conditions which are of most frequent occurrence, *i.e.*, the *mean* conditions, shall be used to

[1] See C. Abbe: "Climatology and its Aims and Methods," *Maryland Weather Service*, I., 1899, 264-304.

characterise it. To give in detail the whole history of the weather phenomena of the district is obviously out of the question. Nevertheless, if we are to present a correct picture, and if the information furnished is to be of practical value, some account should also be given of the extent to which, in individual cases, there may be departures from the average conditions.

Climatology as distinguished from meteorology.—Climatology is but a part of meteorology when the latter term is used in a broad sense. In fact, a sharp line cannot be drawn between these two branches of science. Meteorology, when taken in a restricted sense, seeks to explain the various atmospheric phenomena by known physical laws, and to discover the causes underlying the succession of atmospheric processes. Meteorology is, therefore, essentially theoretical. It dissects this complex body of atmospheric processes in order to correlate the simpler phenomena with the fundamental principles of physics. Climatology is essentially more descriptive in character than meteorology. It aims to present as graphic a picture as possible of the way in which all the atmospheric phenomena work together at any place on the earth's surface. Climatology must treat the different atmospheric processes separately only in so far as this is unavoidable, for it is clear that occurrences which coincide in point of time have to be described one after another. Climatology must give us a mosaic-like picture of the different climates of the world; but it must also present these facts in a systematic way, by grouping together climates which are naturally related. Thus order and uniformity are secured, the mutual interactions of the different climates are made clear, and climatology becomes a scientific branch of learning.

Meteorology, taken in its narrower sense, requires some knowledge of climatology in order to fulfil its work. Climatology, on the other hand, must avail itself of the teachings of theory if it is to solve the scientific problems set before it, which concern the causes for the arrangement of the different climatic groups, and for their influences upon one another. Climatology presupposes a knowledge of the most important teachings of meteorology, just as meteorology presupposes a knowledge of climatic facts. If, therefore, only one of these two subjects is to be treated in a single book, it is merely left for the author, having in mind the special object and the desired size of his book, himself to decide how much of the fact and teaching of the other branch shall be included, and how much taken for granted.

The climatic elements.—The various atmospheric processes and conditions whose interactions determine the climate of any place, are

called *climatic elements* or *factors*. They are temperature, humidity, rain or snow, velocity and direction of the wind, etc. Scientific climatology must endeavour to find numerical expressions for all the climatic elements. Actual measurements are necessary if we are to have statements which shall be strictly comparable, and if we are to gain distinct conceptions of the different meteorological conditions. Instead of such vague statements as "the winter climate of the place is severe"; or, "the summer is windy and changeable," we should have the recorded temperatures and wind velocities, as well as the amount of their variations. Such short, comprehensive, verbal descriptions are admissible only when the numerical values of the climatic elements are also given. Accuracy, and the possibility of comparison, are the first requirements in the description of any climate. These are fulfilled when the numerical values of the individual climatic elements included in such an account were obtained with similar instruments, and according to the same method.

Climatic descriptions gain decidedly in clearness if a certain uniformity of treatment is always adhered to. This is accomplished by discussing the separate climatic elements in a systematic order, based on their relative importance. But as soon as we attempt to distinguish between these elements on the basis of their relative importance in the description of a climate, another difference becomes apparent between the standpoint taken by climatology and by meteorology in their view of atmospheric phenomena. In meteorology, the greatest importance is attached to those phenomena upon which a considerable number of other atmospheric processes depend, and which may therefore be regarded as primary phenomena. In climatology, taken in its restricted sense, those meteorological phenomena become the most prominent which have the greatest influence upon organic life. The importance of the different climatic elements is therefore determined from an outside standpoint.

Definition of climate.—Climatology is thus seen to be a branch of knowledge which is in part subordinate to other sciences and to practical ends. As a matter of fact, this idea is usually incorporated into the first definition of climate, "as the sum total of the meteorological conditions in so far as they affect animal or vegetable life." Humboldt gives the following definition:[1] "The term *climate*, in its broadest sense, implies all the changes in the atmosphere which sensibly affect one's physical condition." This point should be borne in mind when we treat climatology as a science auxiliary to geography.

[1] *Kosmos*, Vol. I., 340.

The term *climate* may, however, be used in a more general way, as was done at the beginning of this introduction, for it seems to the writer perfectly proper to speak of climate as existing on the earth at a time when vegetable and animal life had not yet appeared.

Order of treatment of the climatic elements.—It has just been said that the meteorological elements which are of the greatest importance to animal and vegetable life are given the most prominence in climatic descriptions. The order in which these different elements are to be considered depends, to some extent, upon the advance made in other fields of research which are concerned with the physical conditions of organic life. Atmospheric influences which are now considered of little importance from a climatic point of view, and concerning which there are at present hardly any numerical data, may turn out to be of far-reaching importance; as, for instance, the value of direct insolation, *i.e.*, the heating due to sunshine; or the intensity of the general sky-light. This circumstance affects only the order of arrangement in the descriptive portion of climatology, and does not, in any way, interfere with a scientific presentation of the subject. Furthermore, it plays no part in the treatment of the fundamental portions of climatology, which deal with the factors that control the distribution of climates and their influence upon one another. This portion of our science is based solely upon the teachings of physics, and upon the laws which govern the interrelations of atmospheric phenomena. To investigate these laws is the province of meteorology.

If we wish clearly to define the limits of the various climates, and if our descriptions of the latter are to be comparable, we must first agree upon the various climatic elements on which diversity of climate depends. At the same time, the meaning of these climatic elements must be clearly defined. One of the chief obstacles to the advance of scientific comparative climatology is the confusion which exists on this very point. There is no clear understanding as to just what elements are essential in systematic descriptions of climate, nor as to the best way to discuss these elements, in order that direct comparison of one climate with another may become possible. This confusion is especially noticeable in geographic monographs and text-books, which necessarily make some mention of climatic conditions in their discussion of the physical features of different countries. The same difficulty is also found in writings on hygiene in which the influences of climate are considered. We shall, therefore, at once proceed to consider and explain the most important climatic factors, as well as the methods of

recording them.[1] A tabulation of the climatic elements at Vienna is given in the following pages, and this will serve as a partial explanation of what is stated in the text, and also as a model for climatic tables in general. The tables here given present a compact summary of the most important climatic elements at Vienna.

[1] The following book is to be highly recommended: Hugo Meyer: *Anleitung zur Bearbeitung meteorologischer Beobachtungen für die Klimatologie*, Berlin, Springer, 1891, small 8vo, pp. 187. Frequent reference is made to this book in the paragraphs which follow.

CHAPTER I.

TEMPERATURE.

Importance of temperature as a climatic element.—Temperature is certainly the most important climatic element. Temperature, when used in climatology, means the total effect of the warmth of the air, and also of radiation. When out of doors, plants and animals are always under the influence of the temperature of the atmosphere which surrounds them, and also of the heat due to radiation from the sun, the sky, the surface of the earth, etc. We shall, for the moment, disregard the heat of radiation, and consider first the temperature of the air. From a meteorological point of view, the temperature of a place is simply the measure of the air temperature. This is obtained by means of observations, properly distributed in point of time, of a thermometer freely exposed to the air, but shielded from direct radiation from the sun or from heated bodies in the neighbourhood. The latter condition is of vital importance if we wish to obtain comparable data concerning the temperature of the air at different places. The effects of such radiation vary greatly, and may, even in the same locality, produce temperatures which differ widely from one another; while the true air temperature is found to be comparatively uniform over considerable areas having similar surface features.

It seems to the author of the greatest importance to call attention to the fact that the mean temperatures of stations scattered over a considerable area, *when reduced to the same period of time*, agree so well as to leave no doubt whatever that they accurately represent the air temperature of that district.[1]

[1] See Jelinek: *Anleitung zu meteorologischen Beobachtungen*, 4th Ed., Leipzig, 1893.

Mean daily temperature: how obtained.—The hours at which observations of temperature are made must be properly distributed throughout the day if we wish to obtain the true diurnal temperature, which corresponds to the mean of 24 hourly observations. From such a series of hourly observations we also learn what the highest and lowest temperatures are, although those can evidently be much more easily obtained from the readings of the maximum and minimum thermometers. If the mean is derived from frequent observations made during the daytime only, as is still often the case, the resulting mean is too high, because the temperatures of the cooler portion of the 24 hours do not enter into the result at all. A station whose mean is obtained in this way seems much warmer with reference to other stations than it really is, and erroneous conclusions are therefore drawn concerning its climate. Thus, the mean annual temperature of Rome was given as 16·4° by a seemingly trustworthy Italian authority, while it is really 15·5°, and the mean temperature of Madrid was stated by Secchi to be 15°, while it is actually only 13·5°. The other case, which occurs when a station is reported to have a temperature lower than its actual one, is found much less frequently.

Hours of observation.—It is best to read the thermometer three times daily, namely, early in the morning, in the afternoon, and in the evening. A knowledge of the morning and afternoon temperatures is very important in most climatological investigations, and even an accurate mean temperature cannot replace the lack of this information. The following groups of hours are recommended for observations of temperature: 6 a.m., 2 and 10 p.m.; 7 a.m., 2 and 9 p.m.; 7 a.m., 1 and 9 p.m.; 7 a.m., 1 or 2 p.m., and 10 p.m. The morning observations should not be made later than 7 a.m. The combination, 8 a.m., 2 and 8 p.m., which has unfortunately been quite generally adopted, is not satisfactory, because the mean of $(8 + 2 + 8) \div 3$ is much too high in summer. The mean derived from observations at 7 a.m., 2 and 9 p.m., is also somewhat too high; but the formula (7 a.m. + 2 p.m. + 9 p.m. + 9 p.m.) ÷ 4 gives an excellent mean which differs in summer by only + 0·1° to + 0·2° from the true mean, based upon the twenty-four hourly observations. The means derived from the daily extremes (that is, from the readings of the maximum and minimum thermometers) also give values which are somewhat too high, the difference being about 0·4° in the majority of climates throughout the year. The combination (sunrise + 2 p.m. + 9 p.m.) gives a fairly accurate mean temperature, if the first observation is really made

exactly at sunrise throughout the year. This combination of hours is, however, not an especially desirable one, for the reason that the time of the morning observation varies from day to day. Observation hours which do not vary are always much to be preferred. In the case of stations at which maximum and minimum thermometers are used, and where, furthermore, the readings of the ordinary thermometer have been made at unfavourable hours (*e.g.*, 9 a.m., 9 p.m., or sometimes even at 3 p.m., as has often been done in accordance with an old set of instructions published in England), the most accurate means are obtained by the formula $\frac{1}{2}$(min. + 3 p.m.), or $\frac{1}{4}$(min. + max. + 9 a.m. + 9 p.m.). The formula $\frac{1}{4}$(min. + max. + 8 a.m. + 8 p.m.) gives means which can be used if necessary. In this last grouping, the errors resulting from the combination of (max. + min.) alone, or of (8 a.m. + 8 p.m.) alone, are brought together, and, to some extent, neutralise one another. Sometimes the observations are made at hours which render it impossible to derive comparable means directly from the readings. In such cases, means may still be obtained if a series of hourly observations extending through several years is available for some neighbouring station. These put us in a position to know how far the temperatures at the special hours in question depart from the true diurnal mean at this neighbouring station. It then only remains for us to correct the temperatures at the station whose mean we wish to obtain by the amount of these departures. This method of correcting a mean, however, becomes the more uncertain the greater the necessary correction, for it assumes an absolutely similar diurnal variation of temperature at both stations. This assumption becomes the more arbitrary the greater the differences between the remaining climatic conditions of the two stations.[1]

[1]For further information concerning the derivation of true daily mean temperatures, and the corrections which are necessary in individual cases, the reader is referred to the following:—Schmid: *Lehrbuch der Meteorologie*, Leipzig, 1860, 269 *et sqq.*; H. W. Dove: "Ueber die täglichen Veränderungen der Temperatur der Atmosphäre," *Abhandl. Berlin. Akad.*, 1846 and 1856; H. Wild: *Die Temperturverhältnisse des Russischen Reiches*, St. Petersburg, 1877, I.-LXVIII.; Jelinek: "Die täglichen Aenderungen der Temperatur," *D.W.A.*, XXVII, 1867; W. Köppen: "Tafeln zur Ableitung der Mitteltemperatur," *Repertorium für Meteorologie*, III., No. 7, 1873; G. Hellmann: *Die täglichen Veränderungen der Temperatur in Norddeutschland*, Berlin, 1875; F. Erk: "Die Bestimmung wahrer Tagesmittel der Temperatur," *Abhandl. München. Akad.*, II. Kl., XIV., Part II., 1883; A. M'Adie: *Mean Temperatures and their Corrections in the United States*, prepared under the direction of A. W. Greely, Washington, D.C., 1891, 4to, 45 pp.; W. Ellis: "On the Difference produced in the Mean Temperature derived from Daily Maxima and Minima as

The mean annual temperature expresses most briefly the temperature conditions of the air at any given place. This is really the mean derived from the 365 successive daily means; but, as the mean temperatures of each month are needed in any case, it is agreed that the mean annual temperature shall be derived from the twelve monthly means. The difference between the results obtained in these two ways, which is due to the inequalities in the lengths of the months, amounts to tenths of a degree (C.) only, even in the most severe climates, where the range between summer and winter is greatest. In central Europe, this difference is one of hundredths of a degree (C.) only. As the annual temperature at the same station varies in different years, the observations must be continued for a period of greater or less length if we wish to obtain a mean annual temperature accurate within some definite limit. The length of this period naturally depends upon the amount of the variation in temperature from year to year. If, for example, we wish to obtain a mean annual temperature which shall be accurate within 0·1°, we need about forty years of observations in central Europe, and about sixty years of observations in northeastern Europe. On the other hand, in an equatorial climate, such as that of Batavia, two years of records would (theoretically) suffice.

Mean monthly temperatures.—The annual temperature alone, however, does not suffice to characterise the average temperature conditions of the atmosphere at a given station, because, in most climates, these conditions vary considerably during the year. These annual periodic variations are expressed in the *mean monthly temperatures* (p. 33, columns 1*a* and 1*b*). No attempt should be made to express the normal annual march of temperature by the mean temperatures of successive ten-day or five-day periods, unless, in exceptional cases, there is a very long series of observations for the same station. The monthly means are much less accurate than the annual means. In fact, even when observations which have been made with the greatest care and accuracy are available, it cannot be expected that the monthly means of most places will be known within 0·1°. The mean temperatures of the same month frequently vary very much from year to year. Thus, the means for December, 1879, and December, 1880, in southwestern Germany, differed by 15°. At Washington, D.C., the mean temperature of January, 1880, was 5·5°; that for January, 1881, 2·4°.

dependent on the Time at which the Thermometers are read," *Quart. Journ. Roy. Met. Soc.*, XVI., 1890, 213-218. See also *Indian Met. Memoirs*, V., 1892-1895, 525, and IX., 1895-1897, 638, on the diurnal variation of atmospheric conditions in India.

The mean temperature of July, 1887, at Washington, was 26·9°; and that for July, 1888, 22·8°. At St. Paul, Minn., the mean temperature of January, 1888, was −18·3°; that for January, 1889, −6·6°. The mean temperature of July, 1890, at St. Paul, was 22·2°; and that for July, 1891, 18·7°. At Edmonton, in the Province of Alberta, Canada, January, 1888, had a mean temperature of −21·7°; while January, 1889, had a mean of −5·6°. At Braemar, Scotland, the mean temperature of February, 1895, was 5·8°; and that of February, 1896, was 3·8°. At Greenwich, England, the mean temperature of February, 1855, was 1·6°; and that of February, 1856, 5·6°. In our latitudes, these ranges are about twice as great in the winter months as in the summer. In the case of Vienna, for instance, almost 400 years of observations would be necessary before the mean temperatures of the winter months could be obtained within a limit of 0·1° of accuracy; while in the case of the means of the summer months only 100 years of observations would be needed. For stations in western Siberia, on the other hand, observations would have to be continued for 800 years in the case of the winter months, but for only 100 years in the case of the summer. These figures are given in order that sufficient emphasis may be laid on the fact that monthly means, even when based on long series of observations, are untrustworthy in regions where these same means may vary greatly from year to year. In the climate of Batavia, as contrasted with the cases just cited, only five years of observations are needed to give accurate monthly means. It may be assumed that the mean monthly temperatures based on 20 years of observations in central and eastern Europe are accurate within about 0·4° to 0·6° for the winter, and within about 0·2° to 0·3° for the summer months.[1]

These examples must suffice to give the reader some idea as to the accuracy of mean temperatures, and may enable him to judge for himself how unnecessary and how confusing it is when temperature readings are given in climatological tables to hundredths of a degree.

Mean annual range of temperature.—The difference between the temperatures of the warmest and the coldest months is an important climatic element, namely, the *mean annual range of temperature.* In the vicinity of Vienna, for example, the coldest month (January) has a

[1] The probable error of the mean decreases in proportion to the square roots of the numbers of years of observation. The error for a mean derived from 40 years of observation would therefore diminish in the proportion of $\sqrt{20:40}=0·71$, and would consequently not be decreased by *one-half*, as might have been expected. The probable errors, given above, would, with 40 years of observations, be ±0·3° and ±0·4° for the temperatures of the winter months.

mean temperature of – 1·5° ; the warmest month (July) has a mean temperature of 19·8°. The mean annual range of temperature is therefore 21·3°. At Thorshaven, in the Faröe Islands, March is the coldest month, with a mean of 3° ; July, with a mean of 10·9° is the warmest. The annual range at Thorshaven is therefore only 7·9°. Batavia has a mean temperature of 26·4° in May, and 25·3° in January, which gives a mean annual range of 1·1°. Quito has a January mean of 13·8°; a July mean of 13·3°; an annual range of 0·5°. The mean January temperature of San Diego, Cal., is 12·1° ; that of August, 20·6° ; the range is therefore 8·5°. The mean January temperature of St. Paul, Minn., is – 11·5° ; that of July, 22·2° ; which gives an annual range of 33·7°. In Canada, the mean January temperature at Esquimault, B.C., is 3·7° ; that of July, 15·6° ; which gives a range of 11·9°. At Edmonton the range is 31·9°, the January mean being – 16·1°, and the July mean 15·8°. At Halifax, on the eastern coast, the range decreases again, being 23·2° ; the January mean is – 5·5° ; and the July mean, 17·7°. At Valencia, Ireland, the mean temperature of January, as determined by the 30 years' record, 1871-1900, is 6·9° ; and that of August is 15·3° ; which gives a range of 8·4°. At Aberdeen, Scotland, the January mean is 3·1° ; and the July mean, 14·1° ; which gives a range of 11·0°. At Yarmouth, England, on the eastern coast, the January mean is 2·9° ; and the July and August means are 15·9° ; which gives an annual range of 13·0°. The amount of this annual range of temperature is made the basis of a classification of climates into those of great ranges, or severe climates, and those of small ranges, or temperate climates. Examples of these classes will be given later.

Supan, in 1880, published the first accurate and detailed chart of the annual ranges of temperature over the earth's surface,[1] using lines of equal annual range of temperature, to which he gave the name "isotalantous."[2]

[1] *Zeitschrift für wissenschaftliche Geographie*, I., 1880, 141-156.

[2] The first to construct a chart of this sort was probably F. W. C. Krecke, in Utrecht (*Prov. Utr. Genootsch. v. Kunsten en Wetenschappen*, 1865). He called the lines "isoparallagae," and his chart shows their course over the northern hemisphere, on the polar projection. Keith Johnston, in 1869, also published a chart, on a very small scale, showing lines of equal annual range of temperature (*Proc. Roy. Soc., Edin.*, VI., 1866-69, 561-579). Finally, Buys-Ballot constructed two large charts showing the positions of the "isoparallagae" in both northern and southern hemispheres, taking the difference between January and July as everywhere indicating the amount of the annual range (*Verdeeling der Warmte over de Aarde*, Amsterdam, 1888). This was a step which Supan had already justly censured Johnston for taking, because the extremes of the mean temperatures do not everywhere come in these months.

Wild then prepared, on a large scale, a similar map of the equal annual temperature ranges over the Russian empire, which was published in his *Die Temperaturverhältnisse des Russischen Reiches* (1881). These charts are reproduced, on a small scale, in the *Atlas der Meteorologie*, 1887 (Plate I.). The latest chart of equal annual ranges of temperature for the world, constructed on the basis of the "Challenger" isothermal charts, was published by Connolly in 1894,[1] and is reproduced in Bartholomew's new *Atlas of Meteorology* (Plate 2).

Annual march of temperature.—It is furthermore important to know how the temperature rises from winter to summer, and how it falls again in the autumn, as well as the usual times of occurrence of the maximum and the minimum. In a word, we need to know the character of the annual march of temperature. For climatological purposes, these various points may usually be determined with sufficient accuracy from monthly means based on several years' observations.

The seasons.—In the temperate zones, the year is divided into four seasons, which depend upon the annual march of temperature. These seasons, when considered from a meteorological point of view, are arranged as follows: Northern Hemisphere. Winter: December, January, February. Spring: March, April, May. Summer: June, July, August. Autumn: September, October, November. Southern Hemisphere. Winter: June, July, August. Spring: September, October, November. Summer: December, January, February. Autumn: March, April, May.

The mid-winter month is the coldest; the mid-summer month, the warmest; the mid-spring and mid-autumn months, and, in fact, the seasons of spring and autumn, as a whole, usually have very nearly the mean annual temperature. This statement is naturally true only of the average march of temperature over the greater portion of the temperate zones. Within the tropics, as well as in the polar zones, this subdivison of the year no longer holds good; for there the annual march of temperature is a different one. The mean temperature of the four seasons can therefore be used only in comparing the temperatures of places within the temperate zones. The seasonal means are, however, occasionally preferable, in some respects, to the monthly means. This is due to the fact that the former are more concise when comparisons are to be made, and more trustworthy when we are dealing with means derived from short series of observations. Nevertheless, the seasonal means can never replace the monthly means, which are indispensable in all climatic comparisons that have to do with extended areas of the earth's surface.

Mean diurnal range of temperature: periodic and non-periodic range.—The diurnal period of temperature variation is of importance, as well as the annual period. The amount of the *diurnal range of*

[1] J. L. S. Connolly: "A New Chart of Equal Annual Ranges of Temperature." *Am. Met. Journ.*, X., 1893-4, 505-6.

temperature, or the diurnal amplitude of temperature, is a very noteworthy climatic element, and should be included in every account of a climate which aims to be at all complete. This element is expressed by the difference between the mean temperatures of the warmest and the coldest hours of the day, and is then called the *periodic amplitude*; or, it is expressed by the difference between the mean minima and the mean maxima of the month, obtained from the readings of a maximum and minimum thermometer, or from hourly observations. The latter, known as the *non-periodic amplitude*, is always greater than the former, or *periodic amplitude*, especially in winter. It is therefore desirable to use only one of the two forms of expression when comparing one climate with another, although, when the requisite observations are available, it is of course advisable to include both forms (see Table, p. 33, columns 5 and 6). As a matter of fact, only the non-periodic amplitude is known for most places, because observations made with maximum and minimum thermometers are very much more common than those made every hour during the twenty-four.

It is a well-known fact that the normal variation of temperature during each day has a minimum about sunrise and a maximum about 2 p.m., or, in the summer, somewhat later. In some places, especially on sea coasts, the maximum comes earlier, namely, soon after noon. The *periodic* diurnal range of temperature, therefore, at the equinoxes, and on the average for the year, for instance, is the difference of temperature between 6 a.m. and 2 p.m. It is, however, perfectly clear that the diurnal march of temperature is not by any means always dependent upon the altitude of the sun. Wind, rain, and clouds cause irregular, non-periodic fluctuations, and the maximum and the minimum temperatures often come at times differing greatly from those which would be determined by the sun alone. The maximum and minimum thermometers register the highest and the lowest temperatures which actually occur from day to day, and the mean of the minima thus recorded is the mean of all the lowest temperatures which have occurred at various times. A similar statement is true with respect to the maximum temperatures. When, however, the mean temperatures of the different hours are calculated on the basis of hourly observations made throughout each month, the irregular rises and falls of temperature which may have occurred during any given hour are almost entirely neutralised. Consequently, the lowest temperature is found to come about sunrise, and the highest, shortly after noon. This march of temperature is known as the *normal* or *periodic* march. It is also clear that the difference between the mean temperatures at sunrise and in the afternoon (*i.e.* the *periodic* amplitude) cannot be as great as that between the mean extremes (*i.e.* the *non-periodic* amplitude). Especially is this true of the winter of the temperate and the frigid zones, where the irregular changes in the winds from warm to cold, and *vice versa,* cause very considerable irregularities in the normal march of temperature. In Vienna, for example, the coldest hour of the day in December is 7 a.m., with a mean temperature of $-1{\cdot}6°$; the warmest hour is 2 p.m., with a mean of $0{\cdot}6°$. The periodic diurnal amplitude is therefore $2{\cdot}2°$.

The difference between the mean daily extremes, as recorded by the maximum and minimum thermometers, is, however, 4·7°, or more than twice as great. In the Kara Sea, in latitude 71° N., the coldest hour of the day during the winter of 1882-3 was 9 a.m., with a mean temperature of −22·7°, and the warmest hour was 9 p.m., with a mean of −21·6°. The periodic amplitude was therefore only 1·1°. On the other hand, the difference between the mean daily maxima and minima was 8·7°, or nearly eight times as great, this being the result of the irregular changes in temperature. At Blue Hill Observatory, in December, the mean temperature of the coldest hour, 7 a.m., is −3·5°, and of the warmest hour, 2 p.m., is +0·2°, giving a range of 3·7°. On the other hand, the difference between the maximum and minimum, 3° and −5·4°, is 8·4°.[1]

Mean temperatures of each observation hour.—The normal march of temperature, as controlled by the regular diurnal course of the sun, may be locally subject to considerable irregularities. This is especially the case on sea coasts, owing to the varying land and sea breezes, and in mountainous regions, as the result of the regular occurrence of cold winds at certain hours. The shadows cast by steep mountain sides also interfere with the normal temperature curve. Peculiarities of this sort are of importance in the discussion of the climates of special localities.

As observations are taken but thrice daily at most meteorological stations, namely, in the morning, at noon, and in the evening, the mean temperatures for these periods usually furnish the only information we have concerning the diurnal march of temperature. It is of the greatest importance in all detailed climatic descriptions that the temperature should be expressed, not only by the daily means of the different months, but also by the means of the three observation hours, so that we may know the mean temperature of a given morning, noon, and evening hour, in every month, for any special station (see Table, p. 33, columns 2, 3, and 4). Two places may have the same daily mean temperature, while one of them has a much lower temperature in the morning, and a much higher temperature in the afternoon, than the other. This is actually the case in the sunny winter climate of the southern valleys of the Alps, as contrasted with the dull, rainy, but mild climate of the western coasts of Europe. As the early morning may be spent indoors, the high temperatures of the afternoon, and, to some extent, also of the evening, are of the most importance from a hygienic standpoint. Such a climate as that first referred to may, therefore, be far more beneficial than the second, although they both have the same mean temperature, and even disregarding for the time

[1] See also A. Woeikof: *Die ganze Temperaturschwankung im arktischen und äquatorialen Seeklima*, *M.Z.*, XIV., 1897, 359-361.

being the direct effect of solar radiation, which is a very important element in the former climate.

Irregular variations of temperature. Variability of the monthly means.—The mean temperatures, and the periodic annual and diurnal ranges obtained from them, are insufficient, by themselves, to give an adequate idea of the temperature conditions of any given place. These mean values indicate only the average state of the atmosphere, and it is important, for many reasons, that we should know what departures may be expected in individual cases. When, for example, we find that the mean (normal) January temperature, at Vienna, is $-1{\cdot}7°$, we learn also that this may generally be expected as the mean January temperature in future years. As a matter of fact, of 100 Januaries in Vienna, 33 had a temperature which differed from the mean by 1°, at the most; departures of 1°-2° occurred 28 times; of 2°-3°, 18 times; of 3°-4°, 10 times; of 4°-5°, 8 times; and of 5°-6°, 3 times. In other words, in 33 per cent. of the cases, the January mean was between $-2{\cdot}7°$ and $-0{\cdot}7°$; in 61 per cent. it was between $-3{\cdot}7°$ and $+0{\cdot}3°$; and in 79 per cent. it was between $-4{\cdot}7°$ and $+1{\cdot}3°$. The decrease in the frequency of these departures with the increase in their amount follows the same law[1] in large climatic districts. It is therefore possible to use the mean itself as a basis for determining the frequency of other mean temperatures which differ more or less from this one.

This fact does not, however, make it unnecessary to give also the highest and lowest mean temperatures of any month during a long interval of time. Indeed, these data are extremely desirable. Thus, for example, in the case of Vienna, the temperature of January kept within the limits of $-8{\cdot}3°$ and $+5{\cdot}0°$ during an interval of 100 years; while that of July varied between 17·4° and 24·6°. At St. Petersburg, during 118 years, these limits were $-21{\cdot}5°$ and $-1{\cdot}6°$ for January [mean temperature $-9{\cdot}4°$]; and 14·1° and 23·2° for July [mean temperature 17·7°]. These ranges are much smaller in the case of the mean annual temperatures. Thus, at Vienna, the range in 100 years was between 7·4° and 11·8°; and at St. Petersburg, in 118 years, it was between 1·3° and 6·5°. It will be noticed that, during the winter, the lowest temperatures fall farther below the mean than the highest rise above it. The reverse is true in summer. It is a peculiarity of European climates, that, as a general rule, higher mean temperatures are more common than the lower in winter, and *vice versâ*

[1] In the case of means based on many years of observations, this may be determined by the laws of probability, when the means of the departures from the normal of the given climate are known.

in summer. At Philadelphia, Pa., the range in 111 years was between 9·3° and 16·2°. At Toronto, Canada, the highest mean in 60 years was 8·4°, and the lowest 4·9°. At Greenwich, England, the highest mean annual temperature in 50 years (1841-1890) was 11·1°, and the lowest, 7·9°.

The foregoing figures give the absolute range of the mean temperatures of the individual months. Dove obtained the departures of the mean temperature of the same month in different years from the general mean temperature of that month as derived from the whole series of observations. He then determined the average of the departures from this general mean, without regard to the fact whether they were positive or negative. These latter averages, therefore, represent the mean departures of the monthly temperatures from their average values (see Table, page 33, column 1*c*). They are also of importance in climatology, because they give us some idea as to the number of years of observations which are necessary in order to obtain means of a certain definite accuracy.[1]

[1] These mean departures of the monthly temperatures were called the "mean variability" by Dove. But as *variability* of temperature has another meaning in climatology, it is advisable to call these mean values what they really are; namely, mean departures, or mean anomalies. Fechner has given a very convenient formula for use in determining, from the mean departure, the probable error of the mean derived from a given number (n) of years of observation. His formula is as follows:

$$\text{Probable error } (W) = \frac{1 \cdot 1955}{\sqrt{2n-1}} \times \text{mean departure} = \text{mean departure} \times c.$$

The following table is useful in the determination of the probable error:

$n=$	20	25	30	35	40	50	60	80	100
$c=$	·191	·171	·156	·144	·134	·120	·109	·095	·085

Suppose, for example, that the mean departure of the mean temperature of December in Vienna, based on the records of 50 years, is 2·4° C. The probable error of the 50 year mean is $2 \cdot 4° \times 0 \cdot 120 = \pm 0 \cdot 3°$ C. In other words, the 50 year mean is accurate only within 0·3°. If we wish to know the number of years (n) necessary to give the probable error, W, when the probable error of the mean of n' years is w', we have the following formula:

$$n = n'(w'^2 : w^2).$$

If we wish the probable error to be $\pm 0 \cdot 1°$ C.,

$$n = 100\ n'w'^2$$

and for December, in Vienna, $= 100 \times 50 \times 0 \cdot 09 = 450$ years. It is naturally assumed, in this calculation, that the mean undergoes no progressive secular variation.

The following are some examples of such mean departures, or of the variability of the monthly means :—

MEAN DEPARTURES OF MONTHLY MEAN TEMPERATURES.

	Winter.	Summer.	Mean.
	°	°	°
Interior of North America,	2·54	1·20	1·95
Western Siberia and Ural,	3·02	1·26	2·02
Northern Russia,	3·43	1·61	2·33
Central Russia,	3·09	1·43	2·05
Northern Germany,	2·02	0·93	1·28
Northern Slope of Alps,	2·28	1·06	1·56
Southern Alps,	1·56	1·02	1·25
Dalmatian Islands,	1·30	0·81	1·17
Italy,	1·35	1·00	1·19
England,	1·41	0·95	1·24

In the interior of North America, in the latitude of northern Italy, the temperature of a winter month fluctuates, on the average, more than $2\frac{1}{2}°$ about the mean. In Russia, this fluctuation is nearly $3\frac{1}{2}°$, while in the coast climate of England it is only $1\frac{1}{2}°$. The variability of the monthly means is much smaller in summer, and is then more uniform everywhere. In Batavia, the average departure of the monthly means is but little more than $\frac{1}{4}°$.

The differences between the mean temperatures of the same month in different years are of little importance from a medical point of view, because these differences are separated by a whole year, during which time much greater changes than these occur in the normal course of the annual temperature curve. Organic life is much more closely affected by the irregular changes of temperature which occur during shorter periods, as by the changes within the same month, and especially by those which take place from one day to the next. The value of such changes as these is a gauge of what is known as the variability of temperature. If these changes are slight, the climate is known as an even or a uniform one; if they are considerable, we speak of the climate as variable or changeable as regards its temperature.

Non-periodic mean monthly and mean annual ranges of temperature.—The simplest expression for such changes in temperature within short periods, during which the normal variations in temperature are slight, is the difference between the highest and the lowest temperatures observed during any one month, or the (non-periodic) *monthly*

range (see Table, p. 33, column 9). The monthly extremes should be obtained from the readings of a maximum and minimum thermometer if the value of the monthly range is to be accurate. When, however, observations are taken at least twice a day, once in the early morning and once two or three hours after noon, the monthly range may be obtained with a fair degree of accuracy from these readings, especially in the winter season. The difference between the maximum and the minimum temperatures recorded during a whole year is known as the (non-periodic) *annual range of temperature.* Whenever several years' observations are available for any station, we may determine the means of the various monthly and annual ranges of temperature. This gives the *non-periodic mean monthly* and *mean annual ranges.* In these mean ranges, accidental cases are lost sight of, and in all that concerns the element of variability of temperature they are far better adapted for purposes of comparison of different climates.

Monthly and annual absolute ranges of temperature.—The difference between the maximum and the minimum temperatures observed in the same month during the whole period of observation may also be noted. This is the *absolute monthly range* of temperature for the month in question, In the same way, the difference between the highest and lowest temperatures observed during the whole period of observation represents the absolute annual range of temperature during that period (see Table, p. 33, columns 10 and 11). These differences have but little value unless they are based on very long series of observations. Furthermore, they cannot be used in making comparisons, because all stations have not equally long series of observations, and the absolute ranges naturally increase with the length of time during which the observations are continued. Again, accidental errors of observation affect single data, such as those expressing the maximum and the minimum temperatures which have occurred once during a long period, to so great an extent that such data cannot be advantageously employed in drawing any important conclusions.

The monthly and annual ranges of temperature which have been designated as *non-periodic* are really the sum of both periodic and non-periodic changes of temperature, for these ranges include all the differences of temperature which are actually experienced during a whole month. In the temperate and the frigid zones, the regular variations in temperature from day to night are overshadowed by the greater changes which are due to non-periodic controls—especially the shifts of the winds—and which are most marked in winter. On the other hand, as the latitude decreases, the normal variation of temperature is less and

less under the influence of irregular controls, and the mean monthly ranges of temperature are but slightly greater than the mean differences in temperature between day and night.

The following examples may serve to illustrate more clearly the meaning of mean monthly and mean annual absolute ranges of temperature. The accompanying table gives the absolute maxima and the absolute minima observed at Vienna during the month of January in the ten years, 1881-1890.

VIENNA.—ABSOLUTE EXTREME TEMPERATURES.
JANUARY, 1881-1890.

1881.	1882.	1883.	1884.	1885.	1886.	1887.	1888.	1889.	1890.
° 5·0	° 11·7	° 10·0	° 13·6	° 3·6	° 9·2	° 6·1	° 12·6	° 6·0	° 12·5
–16·4	–6·8	–11·4	–9·7	–14·4	–11·1	–16·0	–14·9	–16·0	–6·4

The mean monthly maximum for January, as determined in this case by only ten years of observation, is 9·0°; the mean minimum, – 12·3°; the absolute mean monthly range, 21·3°. The same method is followed in the case of the other months. The mean of the absolute maximum and the absolute minimum temperatures recorded every year is the absolute mean annual range of temperature. For example:—

CAIRO.—ABSOLUTE ANNUAL EXTREMES OF TEMPERATURE.
1884-1893.

1884.	1885.	1886.	1887.	1888.	1889.	1890.	1891.	1892.	1893.
° 44·8	° 39·6	° 45·2	° 43·3	° 44·3	° 44·2	° 44·0	° 41·6	° 40·8	° 40·6
1·7	5·0	2·6	1·7	2·4	2·9	1·0	2·2	3·8	2·0

The mean of the absolute maxima is 42·9°; the mean of the absolute minima is 2·5°. These are therefore the mean annual absolute extremes of temperature at Cairo, and the corresponding mean annual absolute range is 40·4°. It is very desirable that these mean annual absolute extremes of temperature should be worked out and recorded, because they are very instructive in the discussion of the temperature conditions of any station. The mean monthly minima of January, and the mean monthly maxima of July, cannot be substituted for these annual extremes for the reason that the absolute winter minima do not always come in January[1] nor the summer maxima in July. The

[1] In our climate, the minima usually come between December and February.

January and July extremes are therefore almost always less than the true annual extremes of temperature. Van Bebber has recently published a chart showing the lines of annual extreme ranges of temperature, as determined by many years of observations.[1] The course of the line of equal annual minima of 0° is of special interest, because this isotherm separates those portions of the earth's surface which usually have frost every year from those which are free from frost. In general, the mean and the absolute annual minima are of more importance than the maxima.

Mean diurnal variability of temperature.—There is another way of expressing the variability of the temperature, which from a hygienic point of view is still more useful. This is to determine the differences of temperature from one day to the next for a whole month, and to obtain the mean of these differences. The value thus obtained represents the mean difference of temperature between two successive days in the same month, and the mean of these differences during the same month, for a series of years, is the *normal variability of temperature* for the given station and the given month (see Table, p. 33, column 12). Ten years of observation suffice to give very accurate values. By taking the differences between the successive daily means, we eliminate from the result the effect of the normal diurnal variation of temperature, and therefore emphasise more clearly the non-periodic changes, or disturbances, in the annual march of temperature. In the normal annual march of temperature the change from one day to the next is so slight, even in severe climates, that it is not directly noticeable. In Vienna, with a climate which is half continental in character, the average normal variation in the daily mean temperatures of two successive days is 0·1°-0·2°. In Yakutsk, which is the most exaggerated case of this sort in the world, this variation is 0·3°-0·4°, and even at the maximum, in spring and autumn, it is only 0·5°.

The amount of the daily variation in temperature has already been considered, and it has also been noted that the climatic influence of this element upon man is considerably diminished by the fact that the low temperature of the early morning can easily be avoided by remaining within doors. This is, however, not the case to so great an extent with rapid changes in the mean daily temperatures, in connection with which the physiological effect of the wind, which almost always

[1] W. J. van Bebber: "Die Verteilung der Wärmeextreme über die Erdoberfläche," *Petermann's Mittheilungen*, XXXIX., 1893, 273-276. Three charts of equal mean annual maxima, minima, and extreme ranges of temperature. These are all reproduced in Bartholomew's *Atlas of Meteorology* (Plate 2).

accompanies marked changes in temperature, must also be taken into account. At climatic health-resorts it is particularly desirable that the variability of temperature from one day to the next should be calculated for each of the three observation hours separately, and especially for the afternoon hour.

There are certain other data which are of even greater importance than the mean of the differences between the successive daily means, especially when we wish to make comparisons of two or more climates, and to give a vivid description of the variability of the temperature. These additional data concern the number of times in each month that these differences between the successive daily means reach a given value; for example, the number of times they remain less than 2°; or reach 2°-4°; or 4°-6°, etc. In this connection the rises and the falls of temperature should be given separately, as for example, how often, on the average, a change of – 4° occurs from one day to the next. This method of statement gives adequate emphasis to the considerable changes of temperature which take place from one day to the next, and which, although of infrequent occurrence, are everywhere very important in their consequences.

Other important temperature data.—The most important single temperature data are the highest and the lowest temperatures which usually occur during the year, and in the different months. These are the so-called *mean monthly extremes* and *mean annual extremes*. These extremes are needed for the determination of the monthly and the annual ranges of temperature, which have already been discussed (see Table, p. 33, columns 7 and 8). It is especially important that we have information concerning the lowest temperatures which may usually be expected in winter, and also as to those which have occurred once in a long series of years. The *lowest* temperatures of winter, and not the *mean* winter, or the *mean* summer temperatures, are the determining factors which, in many cases, fix the climatic boundaries of vegetation and of the cultivation of the soil. The highest temperatures of summer are much less important. They are also much more uniformly distributed, and depart less from the mean. The method of obtaining these mean monthly and mean annual extremes has already been explained (see p. 18).

In addition to giving the mean annual minimum temperature it seems to the author very advisable to determine the frequency, or the probability, of the single occurrence of low temperatures of a certain value.

When a certain degree of cold, *e.g.*, – 10°, has occurred once in a given winter, a second occurrence of the same low temperature in the same winter is in many

respects of no particular moment. If, for instance, certain delicate plants have been killed by a severe frost, the recurrence and the duration of the frost are of little consequence. It is therefore very important for the botanist and the farmer to know the probability of the single occurrence of certain freezing temperatures in a given climatic district. This probability cannot be directly determined from the mean annual minimum, as is shown by the following example :—

PROBABILITY OF THE OCCURRENCE OF MINIMUM TEMPERATURES OF -10°, AND BELOW, AT KRAKAU, KLAGENFURT, AND ON THE OBIR.

Station.	Mean Annual Minimum.	Minimum Temperatures. -10°	-15°	-20°	-25°	-30°
	°	Probability of Occurrence.				
Krakau, ...	-21·2	1	0·90	0·63	0·40	0·07
Summit of Obir (2044 m.),	-21·0	1	1·00	0·74	0·27	0·00
Klagenfurt, . . .	-21·7	1	0·90	0·57	0·20	0·03

The probability of very low temperatures is much greater at Krakau than on the Obir and at Klagenfurt, notwithstanding the fact that these stations have virtually the same mean minimum. A temperature of -30° has not yet been observed on the Obir, although it has been noted at Krakau and at Klagenfurt. For purposes of comparison the following figures are also given :—

PROBABILITY OF A MINIMUM TEMPERATURE OF ZERO, AND BELOW, AT VIENNA, MILAN, TRIEST, AND LESINA.

	Minimum Temperatures. 0°	-5°	-10°	-15°	-20°
	Probability of Occurrence.				
Vienna, .	1·00	1·00	0·85	0·55	0·05
Milan, . .	1·00	0·83	0·25	0·04	0·00
Triest, . .	1·00	0·40	0·10	0·00	0·00
Lesina, . .	0·68	0·14	0·00	0·00	0·00

A minimum of -5° occurs every winter in Vienna ; in Milan, eight times in ten years ; in Triest, only four times in ten years ; and in Lesina, once in every seven years.

It has already been pointed out that, owing to the fact that the number of years of observation varies greatly at the different stations, it is not well to give only the highest and the lowest temperatures which have occurred once during the entire period of observation. These data are practically worthless for purposes of comparison, because they are too largely influenced by the chance weather conditions of the different years. They should, therefore, never be given without the

mean monthly and the mean annual extremes, although they are less essential than the latter.

Frequency of occurrence of certain special temperatures.—A statement concerning the frequency of occurrence of certain groups of temperatures (*e.g.*, for every degree or two), is of great value in giving a clear idea of the temperature conditions of any locality. In view, however, of the considerable expenditure of time necessary in making this calculation, and of the amount of space occupied by the tables setting forth the results, such an investigation can be carried out only in single cases, in special monographs on the climate of a given locality. Calculations of this sort, concerning the frequency of certain temperature-groups, have been made by Hugo Meyer, Sprung, Perlewitz, and lately, also by Mazelle.[1]

The theoretical considerations involved in these studies are of special interest, and particularly the explanation of the relation between the data as to frequency and the mean temperature. This matter can be only briefly referred to here. It should, however, be pointed out that these studies make it perfectly clear that the mean temperature does not by any means always coincide with the temperature which is of most frequent occurrence. The mean temperature is therefore not at the same time the most probable temperature (*i.e.*, the temperature which is most likely to occur), although the two do not differ very much. This results from the fact that the departures of the observed temperatures from the mean are not symmetrically distributed on both sides of the mean, the negative departures, for example, being much greater than the positive in the climate of central Europe in winter. A similar fact has already been noted in the case of the departures of the means of the individual months from the corresponding means based on a long series of observations. In summer, the observed

[1] H. Meyer: "Ueber die Häufigkeit des Vorkommens gegebener Temperaturgruppen in Norddeutschland," *M.Z.*, IV., 1887, 428-442; A. Sprung: "Ueber die Häufigkeit beobachteter Lufttemperaturen in ihrer Beziehung zum Mittelwerthe derselben," *M.Z.*, V., 1888, 141-145; P. Perlewitz: "Häufigkeit bestimmter Temperaturen in Berlin," *M.Z.*, V., 1888, 230; E. Mazelle: "Beziehungen zwischen den mittleren und wahrscheinlichsten Werten der Lufttemperatur," *D.W.A.*, LXII., 1895, 57-94; "Beitrag zur Bestimmung des täglichen Ganges der Veränderlichkeit der Lufttemperatur," *S.W.A.*, CIV., 1895, 1015-1082. (This concerns the frequency of the mean daily temperatures and of the mean hourly temperatures in the different months, at Triest and Pola, and also the relation to variability of temperature.) See also C. Chambers: *The Meteorology of the Bombay Presidency*, London, 1878, 47; and J. Hann: "Scheitelwerth und Mittelwerth im tropischen Klima," *M.Z.*, XVI., 1899, 314-315.

temperatures are more symmetrically grouped about the mean. These conditions naturally vary in different climates, but it nevertheless remains true, as a general rule, that the mean temperature does not at the same time always represent the most probable temperature. The studies of Perlewitz show the following conditions for Berlin, as based on the records, for the years 1848-1885.

BERLIN.—DAILY TEMPERATURES, 1848-1885.

	Dec.	Jan.	Feb.	June.	July.	August.
	°	°	°	°	°	°
Mean temperature, . .	0·8	−0·3	1·2	17·5	19·0	18·1
Most frequent temperature, .	1	2	2½	17·0	18	17½
Upper limit, ...	11	11	11	25	29	28
Lower limit, ...	−15	−19	−19	8	10	10

In central Europe, in the winter season, the mean temperatures are somewhat below the most frequent temperature (*i.e.*, of most frequent occurrence), because the extreme temperatures fall farther below the means than they rise above them. In summer, the condition of things is nearly reversed, and the differences are smaller. At Triest, the most frequent daily mean temperature is higher than the monthly mean throughout the year, this difference being on the average about +1°. The most frequent temperatures are those above the mean value, as is the case in central Europe in winter. Hence it follows that the temperatures which are below the mean must fall farther below it than those which are above the mean rise above it. Meyer suggests the term *Scheitelwerth* for the temperature of most frequent or most probable occurrence.[1]

"The *Scheitelwerth*," he says, "is that value about which the individual values group themselves most closely, according to their relative proportions."[2]

If anyone should select, at random, one out of the whole mass of individual values, there would be a greater probability of his happening upon this median value than upon any other.[3]

[1] The term *Scheitelwerth* has no exact English equivalent. "Crest value" is an unsatisfactory rendering. F. Galton calls this the *median value*, and this name is coming into use. See Galton : *Natural Inheritance*, also *Nature*, LXI., 1899-1900, 102-104.

[2] Fechner has, for this reason, called this the *dichtester Werth*, which may be rather unsatisfactorily translated as the *most crowded value*.

[3] H. Meyer : *Anleitung*, pp. 16 *et sqq.*

Meyer urges that these "prevailing" values be worked out for the different observation hours, as well as for each month, and he even goes so far as to prefer the median values to the means. It is very desirable that these median values should be ascertained in all special climatic studies which are based upon long series of observations, and particularly in the case of certain of the meteorological elements ; but the substitution of these values for the means is out of the question.

Duration of certain special temperatures.—Data concerning the time during which certain special temperatures usually prevail are particularly important because of the bearing which such conditions have upon many phases of animal and plant life. When the mean daily temperatures based upon a long series of observations are known for any station throughout the year, it is easy to determine how many days the temperature remains below freezing. This is a climatic element of marked importance, as are the data relating to the number of days when the mean daily temperatures are over 5°, over 10°, over 15°, etc. When, as is usually the case, these accurate daily means are not available, the annual march of temperature is indicated by a curve constructed from the monthly means, and this curve shows when the temperature reaches 0°, and when it leaves 5°, 10°, etc. The duration of these periods is likewise obtained by this same method. In the case of Buda Pest, for example, we have the following conditions:—

DATES AND DURATION OF TEMPERATURE PERIODS AT BUDA PEST.

Temperature Grades, ...	$< 0°$	$> 5°$	$> 10°$	$> 15°$	$> 20°$
Date when reached, ...	Feb. 16	Mar. 20	Apr. 12	May 11	June 17
Date when left, ...	Dec. 5	Nov. 5	Oct. 18	Sept. 24	Aug. 20
Duration of the period, in days, .	74	231	190	137	64

The advance of these different grades of temperature from south to north in the spring, and their return from north to south in the autumn, as well as their movement in a vertical direction, up and down mountain sides, is shown in a very instructive manner by means of such tables. The former case, that is, the seasonal movement of temperatures, is illustrated by means of the instructive charts showing the migration of isotherms in Europe, drawn by Hildebrandsson & Högbom.[1]

[1] H. H. Hildebrandsson: "Marche des Isothermes au Printemps dans le Nord de l'Europe." Read before the Royal Society of Sciences, at Upsala, September 24, 1880. (*Z.f.M.*, XVI., 1881, 340-343.) A. G. Högbom : "Gang der Isothermen

The following table illustrates a case of the second kind.[1]

ADVANCE AND RETREAT OF THE MEAN DAILY TEMPERATURE OF 5° IN THE VICINITY OF VIENNA.

	Station.				
	Mödling.	Schwarzau.	Schneeberg.	Raxalp.	Obir.
Height in meters,	240	620	1460	1820	2050
First noted,	Mar. 21	April 6	May 2	May 29	June 7
Last noted,	Nov. 8	Oct. 28	Oct. 16	Sept. 24	Sept. 24
Duration in days,	233	206	167	118	109

At Innsbruck (600 m.), the temperature of 5° is first observed on March 23; in the valley of Vent (1880 m.), it is not recorded until May 25; and at the mines on the Schneeberg, at Sterzing (2370 m.), not until June 14. By the middle of September it begins its retreat from the latter altitude; reaches Vent at the end of September, and Innsbruck on November 4. In the valley of the Etsch, on the other hand, the same temperature is observed as late as November 20, and at Riva, on the Lake of Garda, even up to December 6. The temperature conditions of a mountainous country are emphasised in a very striking way by means of such data as these.

Supan has published a valuable detailed report, together with a chart, dealing with the duration of the principal thermal periods in Europe;[2] and Tümmler has investigated this subject for the special case of Germany.[3]

Köppen's treatise on the thermal zones of the world, as based upon the duration of hot, temperate, and cold periods, should also be mentioned here.[4]

Temperature control of vegetation. Accumulated temperatures.—In his *Géographie botanique*, de Candolle has attempted to lay down certain

im Herbste im Norden Europas," *Z.f.M.*, XIX., 1884, 112-115 (abstract). See also M. W. Harrington: "The Advent of Spring," *Harper's Monthly Mag.*, LXXXVIII., May, 1894, 874-879.

[1] J. Hann: "Die Temperaturverhältnisse der oesterreichischen Alpenländer," Part III., *S. W. A.*, XCII., 2, 1885. Tables, 56-58.

[2] A. Supan: "Die mittlere Dauer der Haupt-Wärmeperioden in Europa," *Petermann's Mittheilungen*, XXXIII., 1887, 165-172, and Plate 10; with lines of equal duration of frost periods, 0°; warm periods, 10°; and hot periods, 20°.

[3] A. Tümmler: *Mittlere Dauer der Haupt-Wärmeperioden in Deutschland*, Halle, 1892.

[4] W. Köppen: "Die Wärmezonen der Erde, nach der Dauer der heissen, gemässigten und kalten Zeit und nach der Wirkung der Wärme auf die organische Welt betrachtet." With a chart, *M.Z.*, I., 1884, 215-226.

laws which shall express the relations between the temperature of the air and the development of plants. He begins with the assumption that the critical temperature herein concerned is that of 6°; that the only temperatures which have any influence upon the different stages of development of plants are those above this point; and that temperatures below 6° play no part. Every plant needs a certain temperature in order to reach its stages of bearing leaves, blossoms, and fruit. This temperature naturally varies in different plants, but it must be made up of daily means over 6°. The mean daily temperatures obtained in the usual way cannot therefore be directly employed for this purpose. A discussion of the numerous critical examinations which have been made of the actual value of these sums-total of temperature is out of place here, but it certainly cannot be denied that these temperatures are of some importance in the relations between climate and vegetation. The London Meteorological Office has, since 1884, regularly published these accumulated temperatures for the different wheat-producing and grazing districts of England for each week, as well as the whole accumulated temperature (above 6°) since the beginning of the year.[1]

On the basis of such sums-total of temperature, Dr. C. Hart Merriam has shown that the geographical distribution of plants and animals in the United States is determined by the temperature conditions. He concludes that the northward distribution of animals and plants of warmer zones is controlled by the total quantity of heat, *i.e.*, the sum of the daily mean temperatures above 6° (42·8°F.), while the southward distribution of the boreal species is determined by the mean temperature of the hottest part of the year. The latter is expressed, in a general way, by the mean temperature of the six hottest weeks of the year.[2]

Merriam has calculated these sums of temperature for many places,

[1] *Weekly Weather Report.* Strachey has proposed certain rules which are designed to facilitate the calculation of these accumulated temperatures, especially with a view to their being obtained directly from the pentad, or weekly, means of the daily maxima and minima. (*Meteorological Office, Quarterly Weather Report* for 1878, Appendix II., "On the Computation of the Quantity of Heat in Excess of any fixed Base Temperature.") See also "Cumulative Temperature," by R. H. Scott, International Health Exhibition, London, 1880; 8vo, 16 pp.; W. Clowes & Son, London.

[2] C. H. Merriam: "Laws of Temperature Control of the Geographic Distribution of Terrestrial Animals and Plants," *Nat. Geogr. Mag.*, VI., 1894, 229-238. Also: "Life Zones and Crop Zones of the United States," *U.S. Department of Agriculture, Division of Biological Survey, Bulletin* 10, 1898.

and, on this basis, has divided the United States into five zones, which he has represented upon three charts. These zones are:—

I. The *tropical* zone, with a total quantity of heat of 14,500° (26,000°); hottest period, over 26° (78·8°).

II. The *lower austral* zone, with a total quantity of heat of at least 10,000° (18,000°); hottest period over 26° (78·8°).

III. The *upper austral* zone, with a total quantity of heat of at least 6,400° (11,500°); hottest period, below 26° (78·8°).

IV. The *transition* zone, with a total quantity of heat of at least 5,500° (10,000°); hottest period, below 22° (71·6°).

V. The *boreal* zone, accumulated temperature below 5,500° (10,000°); hottest period, below 18° (64·4°).

The determination of *effective* accumulated temperatures is to be recommended for use in detailed climatic monographs, especially those which deal with regions having a variety of local climates, in consequence of differences of altitude, the influence of large bodies of water, etc. These data would be useful as a help towards giving a more graphic account of these local climates, as well as in a study of the practical value of such temperatures for agricultural and botanical purposes. In calculating these temperatures, observations which are absolutely synchronous should alone be used.

Climatic data relating to frost.—*The average date of the last frost* in the spring and of the *first frost* in the autumn is of importance for many reasons, and from these data, the *number of days without frost* can be determined. The *number of days with frost, i.e.*, the number of days on which the temperature falls below freezing, and *the duration of the periods of freezing weather, i.e.*, the number of consecutive days with temperatures below freezing, are also of interest. Additional data of considerable value in a detailed account of the temperature conditions at a station are the *number of winter days, i.e.*, those days on which the temperature does not rise above freezing, even in the afternoon, and *the number of summer days, i.e.*, days on which the afternoon temperature reaches or exceeds 25°.

Summary.—Summarising briefly the results of the preceding discussion, we may say that the elements of the air temperature which are most important in the correct presentation of any climate are the following:—

1. The mean monthly and mean annual temperature of the air.

2. The extent of the mean diurnal range of temperature for each month.

3. The mean temperature at the different observation hours (for

each month), or at least the mean temperatures of an early morning hour, and of an early afternoon hour.

4. The extreme limits of the mean temperatures of the individual months. Also, when there is a long series of observations (*i.e.*, over twenty years), the mean variability of the monthly means.

5. The mean monthly and mean annual extreme temperatures, and the resulting non-periodic mean monthly and mean annual ranges, as well as the mean minimum and mean maximum temperatures for the year.

6. The absolute maximum and minimum-temperatures observed within a given interval of time. The length of this interval should also be noted.

7. The mean variability of temperature as expressed by the mean of the differences between consecutive daily means. This factor is also expressed by the frequency of changes of temperature of a given amount between these daily means, as, for example, changes for intervals of 2 degrees.

8. The average limits, or dates, of frost in spring and autumn, and the number of days free from frost.

The effect of local controls upon air temperature. "City temperatures."—If the air temperatures as recorded by thermometers properly exposed within cities, are compared with the temperature readings obtained simultaneously in the open country near by, it will be noticed that there are differences of greater or less amount between the two sets of observations. As a general rule, it is found that the mean annual temperature of the air in places where there are many buildings is from 0·5° to 1° too high. The differences are greatest in the morning and evening, and least at noon. The diurnal range of temperature is smaller in cities, especially in summer. Renou was the first to demonstrate clearly the difference in temperature between Paris (Observatory) and the surrounding country.[1]

The mean temperature which is usually given for Paris is 0·75° too high; and this is likewise true for Brussels, London, and other cities. The mean temperature of the city of Vienna is 9·7°; that of the surrounding country, 9·2°. The mean temperature of the city of Berlin is 9·1°; that of the surrounding country, 8·6°. The latter reading only should be used in the construction of isothermal charts.

[1] Renou: "Instructions météorologiques," *Annuaire de la Soc. Mét. de France*, III., 1885, 79; "Différences de Température entre Paris et Choisy-le-Roi," *ibid.*, X., 1862, 105-109; "Différences de Température entre la Ville et la Campagne," *ibid.*, XVI., 1868, 83-97. Mahlmann had already pointed this out in 1841, but only in a general way (*Monatsb. der Gesellsch. für Erdkunde*, II., 1841, 55). See also *Z.f.M.*, XX., 1885, 460; and *M.Z.*, XII., 1895, 37-38.

The author has himself collected a number of cases of such differences in temperature between city and surrounding country.[1]

Perlewitz and Hellmann[2] have discussed at length these differences in the case of Berlin. Hellmann also took into consideration the differences in the exposures of the thermometers, and found that Berlin is 0·3° warmer than the surrounding country in winter; 0·6° warmer in spring and summer; and 0·4° warmer in autumn. The evening temperatures in Berlin, however, are 1·2° higher in spring and summer; and 0·8° higher in the mean annual.

In the case of Paris (Tour Saint-Jacques as compared with Parc Saint-Maur) there was found to be a difference of +2·3° on summer nights; the temperatures are the same at noon, and the difference is +1·1° in the diurnal and the annual mean. The city is warmer than the country by these amounts. The mean minima are much higher in cities, while the mean maxima may be the same as those of the country, or sometimes even lower. The cooling by radiation, at night, is much greater in the open than in places which are built up. Eaton calculates that the burning of gas and of coal in London develops sufficient heat to raise the mean temperature of a stratum of air 30 meters thick, over an area of 300 square kilometers (118 sq. miles), 1·2° an hour.[3]

The author has compared the mean temperatures of places within the influence of the Vienna Forest, in the vicinity of Vienna, with those of other stations near by in open country, and with those of the city of Vienna itself. The following results were obtained:—

MEAN TEMPERATURES OF VIENNA CITY, COUNTRY, AND FOREST (1851-1880), ALTITUDE ABOVE SEA LEVEL THE SAME IN ALL CASES.

	January.	April.	July.	October.	Year.
	°	°	°	°	°
Vienna City, .	−1·2	10	20·4	10·5	9·7
Vienna Country,	−1·5	9·6	19·8	10·1	9·2
Vienna Forest, .	−1·5	9·0	19·2	9·6	8·8

[1] J. Hann: "Ueber den Temperaturunterschied zwischen Stadt und Land," *Z.f.M.*, 1885, XX., 457-462; and "Die Temperaturverhältnisse der oesterreichischen Alpenländer," II., *S.W.A.*, XCI., 2, 1885, 425-453; *M.Z.*, II., 1885, 463-465.

[2] *Das Wetter*, VII., 1890, 97-109; and G. Hellmann: *Jahresbericht des Berliner Zweigvereins der Deutschen Meteorologischen Gesellschaft für* 1894, 8.

[3] *Quart. Journ. Roy. Met. Soc.*, III., 1877, 313.

The difference is unimportant in winter, but it reaches 0·6° in summer; and in the interior of the city it is even more than 1°. The following differences of temperature at the several observation hours show how this considerable contrast between city and country comes about:—

VIENNA FOREST COMPARED WITH VIENNA COUNTRY, 1875-1884.[1]

	7 a.m.	2 p.m.	9 p.m.	Mean.
	°	°	°	°
Winter,	−0·8	0·0	−0·8	−0·6
Summer,	−1·1	−0·2	−2·3	−1·4

It is seen that the difference is slight at the warmest hour of the day, while it is considerable in the evening, and probably also at night, especially in summer. The effect of the marked loss of heat, due to nocturnal radiation from a thick cover of vegetation, can plainly be seen in these data. The cooling due to evaporation probably also plays a part. The difference in temperature would have been still greater towards the interior of the city. The cool, damp evenings and nights of wooded areas, as contrasted with the open country, and especially with cities, are clearly indicated in these observations.

The absolute winter minima are much less marked in the interior of cities than in the surrounding open country. For instance, the minimum, in the city of Berlin, in January, 1893, was −23·3°, and in the country, −31·0°; while the absolute summer maxima are hardly any higher in cities, and sometimes even lower, when the thermometers are well exposed and properly sheltered. The temperature which is actually felt in the city, under the influence of the radiation from the heated walls of buildings and the reflection from the bare ground, is, however, very different from that felt in the country.[2]

The only discussion of this subject for the United States is that by Mendenhall,[3] who points out that during the cold waves of January, 1884, the mean minimum temperature registered at the regular stations of the United States Signal Service in Toledo, Cleveland, Columbus, and Cincinnati, Ohio, were from 1·7° to 8·3° higher than

[1] Hadersdorf as compared with the Meteorological Observatory. J. Hann: "Die Temperaturverhältnisse der oesterreichischen Alpenländer," II., *S. W.A.*, XCI., 2, 1885, 403-453.

[2] See also J. Hann: "Temperatur von Graz Stadt und Graz Land," *M.Z.*, XV., 1898, 394-400.

[3] T. C. Mendenhall: "A Question of Exposure," *Science*, III., 1884, 306-308.

those recorded at voluntary observers' stations outside of large cities. The minimum temperatures recorded in the large cities were, on different days, 2·8°, 4·4°, 7·2°, 10·6°, and 15° higher than those noted at stations of the State Weather Service. Mendenhall concludes that for measurements of temperature it would be well to put stations *near*, rather than *in*, large cities, and at sufficient distance from them to be free from purely local conditions.

Importance of mean temperatures derived from observations made during the same periods of time.—In determining the differences in temperature between neighbouring places, the greatest degree of accuracy is attained when the differences are based on observations made during the same years, and especially when the hours of observation are the same. Differences of temperature can be determined with much greater accuracy when the observations are made at the same hours, if only for three or four years, than when many years' records are available and the hours are not the same. Thus, for example, the winter mean of Augsburg is – 1·3°, based on observations made between 1812 and 1878; that for Munich, based on records made between 1825 and 1856, is – 1·8°. Thus Munich seems to be 0·5° colder than Augsburg, which in itself is not likely. As a matter of fact, the more recent synchronous observations (1879-1881) show that Augsburg is somewhat colder than Munich in winter, although the difference amounts to only 0·1°. Augsburg is 0·1° warmer than Munich in summer, and the mean annual temperatures are the same. It is absolutely essential, in view of the very great differences in the temperatures of individual years, especially in winter, that means derived from the same years should alone be employed in climatic investigations; or else, that these means should be reduced to the same period. The December mean for Krems, situated on the Danube, 55 kilometers from Vienna, was, for example, 1·5° for the five years 1880-1884. If it were desired to make a comparison of Vienna with Krems, and there were available for Vienna only the mean derived from observations made during the years 1879-1883 (*i.e.*, – 0·2°), which, it will be noted, is a period differing by only one year from that available for Krems, it would be supposed that Krems is nearly 2° warmer than Vienna. The error would, in this case, be very considerable. If, on the other hand, we take the differences between the December means of Krems and of Vienna, in those four years which correspond in the two series, *i.e.*, 1880-1883, we find them to be 0·2°, – 0·5°, – 0·3°, – 0·2°, which gives a mean of – 0·2°. Applying this mean departure to the 50-year December mean for Vienna (Observatory), which is – 0·4°, we obtain, as

TABLE I.

CLIMATIC DATA FOR VIENNA, LAT. 48° 12′ N., LONG. 16° 22′ E., ALTITUDE ABOVE SEA LEVEL, 194 METERS.

A.—Temperature (in degrees Centigrade).

	Mean derived from 24-hour observations (20 years).	100-year Mean (reduced)	Mean departures of the Means (90 yrs.).	Means for three observation hours (20 years).			Diurnal range of temperature (20 years).		Mean monthly and annual Extremes (20 years).		Mean monthly and annual Range.	Absolute Extremes, 1829-1875.		Mean diurnal variability of temperature.
				6 a.m.	2 p.m.	10 p.m.	Periodic.	Non-periodic.						
	°	°	°	°	°	°	°	°	°	°	°	°	°	°
December, . .	−0·8	−0·3	2·3	−1·5	0·6	−1·0	2·1	4·7	9·6	−11·2	20·8	19·1	−22·6	2·0
January, . .	−1·3	−1·7	2·5	−2·3	0·3	−1·6	2·7	4·9	9·7	−12·1	21·8	18·8	−25·5	2·1
February, . .	0·4	0·1	2·2	−1·2	2·6	0·1	3·8	6·1	11·4	−10·0	21·4	20·0	−20·0	2·0
March, . . .	4·2	4·3	1·8	1·6	7·4	3·6	5·9	7·8	16·7	−5·9	22·6	24·3	−13·3	1·8
April, . . .	10·0	9·9	1·7	6·2	14·0	9·0	7·8	9·6	23·9	−1·0	24·9	28·8	−7·0	1·9
May, . . .	15·1	15·1	1·5	11·4	19·3	13·8	8·2	10·2	28·5	2·7	25·8	36·0	−1·6	1·8
June, . . .	18·6	18·8	1·2	15·5	22·4	17·1	7·6	9·9	31·5	9·1	22·4	37·8	3·8	1·9
July, . . .	20·3	20·5	1·3	16·9	24·3	18·9	7·9	10·1	32·6	11·0	21·6	38·8	8·0	1·9
August, . .	19·6	19·7	1·3	16·0	23·7	18·2	7·9	9·7	32·9	9·8	23·1	37·5	5·6	1·8
September, . .	16·1	15·9	1·2	12·2	20·4	14·8	8·2	9·6	28·3	4·9	23·4	33·5	−0·6	1·7
October, . .	10·5	10·0	1·4	7·7	14·3	9·5	6·6	8·3	23·2	0·6	22·6	27·1	−6·8	1·5
November, . .	3·7	3·9	1·4	2·5	5·5	3·3	3·1	4·9	14·9	−5·9	20·8	21·3	−15·0	1·8
Year, . . .	9·7	9·7	0·74	7·1	12·9	8·8	5·9	8·0	33·9	−15·1	49·0	38·8 July 14, 1832	−25·5 Jan. 22 1850	1·9
Column, . .	1a	1b	1c	2	3	4	5	6	7	8	9	10	11	12

the reduced 50-year mean for December, at Krems, the value $-0{\cdot}6$. This method of reduction gives results of great accuracy, and makes it easy to obtain monthly means comparable within $\pm 0{\cdot}1°$. Such accuracy is necessary in studies of local climates. The variability of synchronous differences of temperature is about ten times smaller than that of the mean temperatures themselves. Therefore, by the laws of probability, the mean difference of temperature between two places which are not very far apart, when based on ten corresponding years of observations, is just as accurate as a 100-year average, based on the mean temperature of these stations.

Radiant heat.—Meteorology is concerned with air temperature alone, and not with radiant heat, except in so far as the temperature of the air is thereby modified. In climatology, on the other hand, radiant heat is of the greatest importance, quite independently of the temperature of the air. In the problems with which meteorology has to deal, such as the causes and the effects of atmospheric movements, the temperature of the air alone plays a part. On this account, our thermometers are protected, so far as possible, from the effects of radiation, because the latter prevents our ascertaining the true air temperature. Direct solar radiation is, however, just as important as the temperature of the air for organic life on the earth's surface, and even for some of the changes which take place in inorganic bodies. It would, therefore, be of the highest importance in climatology if there were available regular and continuous measurements of the energy and the amount of the solar radiation which actually reaches the earth's surface in different climates. Unfortunately, there are as yet no such measurements. In fact, we hardly have the means, at the present time, for undertaking them.

Nature and effects of solar radiation.—The energy of the direct solar rays can be measured only by their effects. These are chiefly of three kinds, namely, optical, thermal, and chemical. For this reason, we speak of *light*-rays, *heat*-rays, and *chemical*, or *actinic*-rays, but it must not be forgotten that these different effects are not the result of corresponding differences in the quality of the rays themselves, but are dependent on the nature of the surface upon which they fall. It is true that solar radiation is of a composite character, but the components differ from one another only in their wave-length, or their period of undulation. These differences can be easily seen in the varying degrees of refraction, depending upon their different wave-lengths, which the solar rays suffer when a beam of them is split in passing through a prism. The "light-waves," using the word *light* to

express radiant energy in general, which are sent out by the sun, are a vibration of the ether.[1]

The different sets of rays possess different degrees of energy, but are otherwise similar in their nature. Their effects, whether they cause an increase in temperature, or bring about chemical decomposition, depend solely upon the nature of the substances which absorb the rays. The rays must encounter some obstruction which interferes with their movement before they can appear in one or another of these three ways. The work which is done by the active energy of the waves in the ether, becomes apparent either as heat, which is molecular motion, or as chemical action, in which the motion has brought about a rearrangement of the atoms. There are no rays which can produce only heat, and there are none which can cause only chemical action. The same ray which, falling upon a thermometer, or upon a sensitive thermopile, demonstrates its existence by the evolution of heat, will, if it falls upon another substance of a certain composition, reveal itself in chemical rearrangements. When we speak of "radiant heat," or of "chemical," or "actinic" rays, we are confusing cause and effect, for we are then putting something into the nature of these rays which does not actually appear until they come in contact with a surface of some particular kind. It is therefore best, in order to avoid misunderstanding, to speak only of *radiant energy.*

The fact that rays of a given wave-length produce unusually active chemical changes in bodies of certain compositions, is not in contradiction to the statement just made. It is this very fact that has led to the use of the term *chemical rays.* Nevertheless, when these very same rays fall upon a sensitive thermopile, or upon the bolometer, they produce heat, and therefore appear as "heat" rays. It must also be noted that recent investigations made with diffraction spectra, instead of with prismatic spectra produced by refraction, have shown that the maximum heating effect in the spectrum lies between the Fraunhofer lines D and E, and not in the infra-reds, as was formerly supposed. The maximum energy of solar radiation is in the yellow; the maximum light effect and the maximum thermal action coincide.

Effect of different rays upon vegetation.—Recent investigations have also shown that the components of radiation which are of most importance in the growth and development of plants are not at the blue and violet end of the spectrum, but toward the red end of the visible

[1] The energy, or the active force, of a wave of given wave-length and intensity is proportional to the quotient of the square of the amplitude divided by the square of the wave-length.

spectrum. In other words, the rays most useful to plants are luminous rays, and are not the so-called "chemical," *i.e.*, substantially the ultra-violet, rays, as was formerly supposed.[1]

In some experiments carried on by William Siemens in London, in the winters of 1879-80 and 1880-81, several plants were fully developed and even brought to the point of bearing fruit, solely by artificial light. Electric lights were used, and the temperature was kept at about 15°. It was found that the electric light, in which the so-called "chemical" rays are very abundant, had to be passed through shades which partially absorbed the blue rays, before it would produce strong and healthy plants. More recently, Flammarion has studied the development of vegetation under the influence of rays of different degrees of refrangibility. The radiation employed was almost monochromatic, and the plants were in all other respects kept, so far as possible, under the same conditions. The growth of *mimosa pudica* may be cited as an illustration. The seed was planted in May.

EFFECT OF DIFFERENT KINDS OF LIGHT UPON THE GROWTH OF *MIMOSA PUDICA*.

Kind of light, .	Red.	Green.	White.	Blue.
		Height of Plants.		
September 6, .	0·22 m.	0·09 m.	0·04 m.	0·03 m.
October 22, .	0·42 m.	0·15 m.	0·10 m.	0·03 m.

The plants attained their greatest height and greatest luxuriance under the influence of the red light. With reference to the amount of development, the different kinds of light were arranged in the series, red, white, green, blue. The red and the yellow rays are the most active in promoting the respiration and the transpiration of the leaves, and the assimilation of carbonic acid.[2]

Griffiths also found that the rays which are most favourable to the active absorption of mineral matter through the roots are identical with those which promote assimilation in the green portions of plants. These he found to be the yellow rays.[3]

[1] Rays of slight refrangibility ("light" and "heat" rays) are most effective in the processes of assimilation in green plants, *i.e.*, in the production of organic matter. The rays of great refrangibility ("chemical" rays) hinder growth, but are said by Sachs to be of the greatest importance in the production of the materials necessary for the development of blossoms.

[2] C. Flammarion: "Etude de l'Action des diverses Radiations du Spectre solaire sur la Végétation," *Comptes rendus*, CXXI., 1895, 957-960.

[3] Griffiths: "Investigations on the Influence of Certain Rays of the Solar

Importance of diffuse daylight for vegetation.—The important part played by diffuse daylight in relation to plants has recently been clearly pointed out by Wiesner. All vegetation which is growing well or luxuriantly is, above all else, dependent upon diffuse light, as well as upon sunlight the intensity of which has been somewhat diminished. The exposure of the leaves is usually determined by the strongest diffuse light at the spot where the leaves are. It is the diffuse light which pours over all portions of the plant equally, and which benefits them all. Wiesner has also made careful measurements of the intensity of this diffuse light in the shadows of different plants and trees. It is interesting to note the discovery that plants need less light as the temperature rises, and more light as it falls. This is true not only for changes of temperature with latitude, but also for those due to differences of altitude. Thus, for example, the minimum of light necessary for *poa annua* at the beginning of March at Cairo produces about 53 calories; while at Vienna it gives 109. The mean temperatures are 15·5° and 2°, respectively (3·5° at noon). When the altitude of the sun is the same at Cairo and at Vienna, 53·3° at noon,[1] the minimum of light at Vienna gives 92 calories, and the air temperature is 10·4°. Even when the sun stands at the same altitude, *poa annua* needs more light in Vienna than in Cairo, because the corresponding air temperature is lower in Vienna.[2]

Photometric and other related observations.—There are certain additional data which would be most important contributions toward giving us a more complete knowledge of the chief climatic elements. These are measurements of the strength of the sunlight, *i.e.*, photometric measurements, and of the heating effects of direct solar radiation, which should be supplemented by measurements of the chemical activity of this radiation. In the sections on solar climate, reference will be made to the few measurements of this kind which are now available. Further than that we can but emphasise once more the importance of these climatic elements which have thus far been so sadly neglected. It is to be hoped that when the seriousness of this deficiency becomes known, the difficulties that still stand in the way of such measurements will speedily be overcome.

Measurements of radiant heat: Actinometry.—In determining the

Spectrum on Root-Absorption and on the Growth of Plants," *Proc. Roy. Soc.*, Edin., XIV., 1886-87, 125-129.

[1] March 3, at Cairo; April 20, at Vienna.

[2] J. Wiesner: "Photometrische Untersuchungen auf pflanzen-physiologischem Gebiete," *S.W.A.*, CII., 1, 1893, 291-350; and CIV., 1, 1895, 605-711.

value of health resorts with reference to their greater or less beneficial effects upon the sick or infirm, it has long been known that, in addition to the warmth of the air, the energy of direct insolation is of great importance as a climatic factor. Indeed, in times of calm, the air temperature itself may be quite unimportant. An unobstructed exposure to solar radiation, combined with a calm condition of the atmosphere, makes it possible for certain high valleys in the Alps, which are especially well sheltered, to rival many southern winter resorts, notwithstanding the extreme cold of the Alpine winters.

"The summer climate of Davos is very similar to that of Pontresina and St. Moritz, in the neighbouring valley of the Engadine—cool, and rather windy; but so soon as the Prättigau and surrounding mountains become thickly, and, for the winter, permanently covered with snow, which usually happens in November, a new set of conditions come into play, and the winter climate becomes exceedingly remarkable. The sky is, as a rule, cloudless, or nearly so; and as the solar rays, though very powerful, are incompetent to melt the snow, they have little effect upon the temperature, either of the valley or its enclosing mountains [mean temperatures: Dec., $-5{\cdot}5°$; Jan., $-7{\cdot}4°$; Feb., $-4{\cdot}2°$]; consequently there are no currents of heated air; and, as the valley is well sheltered from more general atmospheric movements, an almost uniform calm prevails until the snow melts in the spring."

Even when the temperatures in the early morning are from $-15°$ to $-20°$, the invalids at Davos go out to walk soon after sunrise (9-10 a.m.), without any special wraps, and many of them even without overcoats. The sky is dark blue, and in the strong sunshine one feels comfortably warm when sitting in front of the hotel with a thin coat on. The rarer atmosphere conducts less heat away from the body, even apart from the effect of the calm. Furthermore, there are no minute watery particles in the air to make it damp and chilly as at lower altitudes. In addition to this, the direct radiation from the sun, together with the heat due to reflection from the snow, give a temperature in mid-winter which is occasionally almost unbearable when one is sitting still in the sunshine.[1]

Actinometers.—Frequent attempts have been made in different places to measure the intensity of radiant heat, although, be it said, the apparatus used was inadequate. Almost the only instrument which is now employed in the systematic record of the intensity of solar radiation, but one that cannot be recommended for accuracy, is the so-called black-bulb thermometer *in vacuo*, which is quite generally used, especially at English meteorological stations. At Davos, in December, this instrument records an average temperature of 39°, with a mean maximum air temperature of $-1{\cdot}5°$, and the black-bulb *in vacuo* may even rise above 62°. This instrument is very much affected by local conditions.

[1] E. Frankland: "Note on some Winter Thermometric Observations in the Alps," *Proc. Roy. Soc.*, London, XXII., 1873-74, 317-328.

The Arago-Davy actinometer consists of a black-bulb thermometer of this sort, and of a second thermometer, similarly adjusted, whose bulb is left bright. These instruments give only relative values, and their readings cannot be used for the determination of the absolute value of the intensity of solar radiation.[1] Crova's actinograph is a valuable instrument for the registration of direct solar radiation, and Angström's electrical compensation pyrheliometer leaves little to be desired for absolute measurements of the same quantity.

"Temperatures in the sun."—The readings of insolated thermometers in a vacuum are, nevertheless, far preferable to the very vague data obtained when an ordinary, or even a blackened, thermometer is freely exposed to the sun. Temperatures obtained by this latter method, and spoken of as "temperatures in the sun," have no scientific value whatever; they cannot be compared with one another, nor can they be used in climatological investigations.

Even with the same intensity of solar radiation, "temperatures in the sun" differing widely from one another, will be obtained with different thermometers, although the instruments are all simultaneously affected by the various, and never wholly avoidable, influences of radiation from surrounding objects. Lamont has already shown that a thermometer with a small bulb, freely exposed to the sun, indicates the air temperature very closely. Therefore it must lose enough heat by reflection and radiation to give it the same temperature as that recorded by a thermometer in the shade. Professor Kunze, of Tharandt, has recently published some similar results, although he obtained greater differences.[2]

The "temperature in the sun," as contrasted with the shade temperature, is, furthermore, a very indefinite factor, because it depends upon the nature of the body which is exposed to the sun's rays. The temperature of a body which is exposed to radiant heat is determined by its power to absorb the rays which fall upon it. This power of absorption, again, varies according as the upper surface of the body is rough or smooth; a thin covering of some foreign material may alter it completely. As soon as the temperature of a body under the influence of radiation rises above that of neighbouring bodies, it loses heat again by conduction and radiation, and a stationary temperature is reached when the amount of

[1] See what Ferrell says on this subject in his papers, "On the Conditions determining Temperature," *Bull., Philosoph. Soc.*, Washington, V., 1883, 91-97, and on "Temperature of the Atmosphere and Earth's Surface," *Professional Papers of the U.S. Signal Service*, No. XIII., 1884, [*Z.f.M.*, XIX., 1884, 500-501.] Also, Maurer: "Zur Theorie des Actinometers Arago-Davy," *Z.f.M.*, XX., 1885, 18-20, and C. Abbe: "Temperature in the Sunshine," *Monthly Weather Rev.*, XXVII., 1899, 310-311.

[2] M. Kunze: "Temperaturangaben eines im Freien aufgehängten Thermometers," *Z.f.M.*, XVII., 1882, 291; W. Köppen: "Studien über die Bestimmung der Lufttemperatur und des Luftdrucks; I. Abhandlung: Untersuchungen über die Bestimmung der Lufttemperatur," *Aus dem Archiv der deutschen Seewarte*, X., 1887, No. 2, Hamburg, 1888; (*M.Z.*, VII., 1890 [33-35]). See also H. A. Hazen: "Thermometer Exposure," *Professional Papers of the Signal Service*, No. XVIII.; "Thermometer Exposure," *Amer. Met. Journ.*, III., 1886-87, 82-92.

heat thus lost just equals the amount gained by the absorption of radiant energy. In addition, there comes into play, as a third factor, the loss of heat resulting from the movement of the air, *i.e.*, the wind. Thus the temperature of a body exposed to the sun's rays depends upon a variety of attendant circumstances, which make the simple statement of the result of no value whatever. The black-bulb thermometer *in vacuo* has the great advantage that the surrounding enclosure of glass almost completely shuts off the "dark" radiation from surrounding objects. This latter radiation varies very greatly in different localities, and is one of the chief obstacles in the way of securing comparable observations of this sort. On the other hand, the luminous rays pass through the glass without suffering any loss to speak of. Furthermore, the conduction of heat by means of wind is entirely prevented. For these reasons, black-bulb thermometers *in vacuo* give results which are fairly comparable. For information concerning the methods used in the measurements of the intensity of solar radiation, the reader is referred to the text-books of physics. A few of the results of such measurements will be given in the section on solar climate.

Reflected heat.—The heat due to reflection from objects on the earth's surface is a climatic element which may locally attain considerable importance, not only because of its effect upon vegetation, the ripening of fruits,[1] etc., but also on account of its beneficial influence upon persons who are weak and diseased, but who may yet remain out of doors. It has long been known, and has lately been directly proved by Dufour, that the reflection from water surfaces is a very considerable source of heat for the hillsides which surround these bodies of water. This heat from reflection has an effect upon the ripening of grapes around the Lake of Geneva and along the Rhine, which is by no means to be undervalued. The reflection and radiation from the mountain sides raise the temperature of the air in valleys, and, to a still greater degree, increase the feeling of warmth.[2]

Dufour obtained the following results in connection with the ratios of direct and reflected heat around the Lake of Geneva:—

RATIOS OF DIRECT AND REFLECTED HEAT ON THE LAKE OF GENEVA.

Altitude of sun, ...	±4°	7°	16°
Reflected heat in percentage of direct heat,	68 %	40-50 %	20-30 %

[1] A. E. Frye reports (in Ms.) that in some parts of California, as *e.g.*, the district of the Santa Ana River south of the San Bernardino Mountains, the glare of the sun from the gravelly and sandy "washes" along the river is markedly stronger on the slopes north of the washes than south of them, and that this affects the oranges grown on the former slopes. As a rule the best orange lands are the alluvial fans or slopes along the southern base of the San Bernardino and Sierra Madre ranges.

[2] The temperature of mountain sides is higher by day than the temperature at the same altitude above the earth's surface.

The reflected portion is thus seen to play the chief part when the sun is low, *i.e.*, in the morning and in the winter of high latitudes. A southern exposure at a considerable height above an extended surface of water, or even of snow, gains the greatest advantage from this secondary source of heat, and is therefore the most highly to be recommended for house sites.

Frankland publishes the following observations upon the effect of heat reflected from surrounding objects.[1] The thermometer exposed to the rays was put upon a background of white paper. An observation at Pontresina gave the following results: 10 feet from a white wall, 38·7°; over an adjoining meadow, 27·7°, or 11° less. At Alum Bay, Isle of Wight, under the influence of the direct radiation and of the radiation reflected from the water, 31·2°; direct radiation only, 25·7°; Lake of Zurich, temperature in direct and reflected radiation, 34°; one mile from the lake, in direct radiation only, 31·5°. As our feeling of warmth and our comfort, when we are out of doors, depend upon the combined effect of direct and reflected radiation, it is clear how important is the influence which the surroundings of a place of residence have upon what might be called the "climatic temperature."

Terrestrial radiation: Nocturnal cooling.—There is another, and a contrasted effect of the loss of heat by radiation which is of great importance climatically, and may be directly observed with much greater ease. This is the nocturnal cooling of the free surfaces of bodies to a temperature below that of the air. On clear nights the temperature of the surface of the earth, or of plants, often falls considerably below that of the air at some distance above the earth's surface. The temperature of the air being that of which we wish to obtain a record, thermometers are protected from the effects of nocturnal radiation by means of shelters. This is necessary because thermometers, like almost all other bodies, are much better radiators than the air itself, which cools but slightly by radiation. Different bodies cool, as the result of nocturnal radiation, by different amounts, as is shown by the varying quantities of dew which form upon their surfaces. For climatological purposes the intensity of nocturnal radiation is best measured by means of a minimum thermometer laid directly upon a surface of short grass, and by means of a thermometer laid on the bare ground and lightly covered with earth.[2]

[1] E. Frankland, "Climate in Town and Country," *Nature*, XXVI., 1882, 380-383.

[2] In climates where snow falls in winter, the thermometers should be laid

The difference between the minimum temperature in the free air and that of the air close to the grass or the surface of the earth, is a measure of the loss of heat by nocturnal radiation. Observations of this sort, although easily made, are nevertheless not available for many climates. The English meteorological stations alone are generally provided with radiation thermometers.

In Vienna, the readings of a minimum thermometer which was freely exposed on the grass averaged lower than those of the minimum thermometer in the shelter, four or five feet above the surface, by the following amounts:—in spring, 1·3°; in summer, 1·8°; in autumn, 1·3°; mean monthly extremes, in spring, 2·1°. We may therefore conclude that frost can occur in the neighbourhood of Vienna even when the mean nocturnal minimum temperature is +2° to +3°. These differences are still greater in drier climates, especially at greater altitudes above sea level; and frost can occur when the air temperature is 5°-6°, if radiation is favoured by a clear sky, and if the absence of wind makes it possible for considerable differences of temperature to be produced between bodies in the air and the air itself. On the dry plateau of Yemen, with a nocturnal minimum of only +8°, Glaser saw the pools in the vicinity frozen over in the early morning.

Henri Dufour has tried to show, in a very instructive way, the effect of different surface colours upon different amounts of warming by insolation and of cooling by nocturnal radiation. In this investigation he used four minimum thermometers, the bulb of one of which he covered with black flannel; that of another, with red; that of a third, with white; while that of the fourth he left bright. These thermometers were exposed to radiation on February 20, 1895, with the following results:—

Thermometer, .	Black.	Red.	White.	Bright.
	°	°	°	°
In the sun, . .	39·5	29·0	23·6	22·0
5 p.m., . . .	5·5	4·5	3·6	1·8
6.15 p.m., . .	−4·5	−5·0	−5·0	−6·0
8 a.m., minimum, .	−10·5	−11·0	−10·0	−10·0

The night on which these observations were made was clear. The colour is seen to have no effect upon nocturnal radiation, as has already been shown by other observations.[1]

Soil temperatures.—Finally, the temperature of the ground is also a climatic element worthy of note. For climatological purposes, observations in the upper portion of the ground, from one to two meters deep, and on the surface itself, suffice. In districts which are seldom visited, or in places where, for other reasons, observations of temperature

directly upon the surface of the snow, and upon the ground from which the snow has been swept away.

[1] *Bibliogr. universelle, Archives des Sci. phys. et nat.*, XXXIII., 1895, 477-478.

extending through a whole year are not to be expected in the near future, the determination of the temperature of the soil at depths of from ten to twenty meters will, if the percolation of water is prevented, serve fairly well as a means of estimating the mean annual temperature. At stations near the equator the depths need be but one meter or more. As the daily variations in temperature hardly extend one meter into the ground, one observation a day at greater depths than this suffices to give good means. The annual means of soil temperatures at a depth of one meter are about 1° higher than those of the air temperature in middle and higher latitudes. From that point on, the temperature increases, as a rule, regularly toward the interior of the earth, on the average about 1° in 30 meters.[1]

Sensible temperatures.—Before leaving the subject of temperature as a climatic element some mention should be made of the conditions upon which our sensation of heat and cold depends.[2] The temperature which we actually experience does not depend solely upon the air temperature, *i.e.*, the reading of the dry-bulb thermometer, but also to a considerable extent upon other meteorological conditions which prevail at the same time, especially upon the amount of air movement. Everyone knows that severe cold may be easily endured if the air is calm, but may become unbearable when there is more movement of the air. On bright, cold, winter days, the springing up of a gentle breeze is sufficient at once to change a temperature which seemed agreeable before into one which is decidedly chilly, and yet the thermometer reading shows no change whatever. Furthermore, insolation and the humidity of the air have a considerable influence upon the temperature which we feel.[3]

[1] The results of eight years' observations of underground temperatures at Oxford, England, are given in *Results of Meteorological Observations made at the Radcliffe Observatory, Oxford, in the Eight Years 1892-99* (Oxford, 1901). An interesting case of the slow penetration into the ground of the high temperatures of a hot wave in South Australia is noted in Sir Charles Todd's *Rainfall in South Australia and the Northern Territory during 1897* (Adelaide, 1900). See *Science*, N.S., Vol. XII., 1900, 851-852.

[2] See also W. J. van Bebber: *Hygienische Meteorologie*, Stuttgart, 1895, 124-145.

[3] Vincent undertook an extended series of investigations in order to show the relation between the temperature of the exposed surface of the skin (T); the air temperature (t); the excess of the temperature as indicated by the actinometer over that of the air, in degrees (d); and the velocity of the wind in meters per second (v). (J. Vincent: *La Détermination de la Température climatologique*, Bruxelles, 1890. Reprinted from *Annuaire de l'Observatoire royal pour 1890*). The observations of air temperature between 6° and 26°, and in calms, showed

In hot climates, and also in the summer of middle and higher latitudes, when the body is usually covered with perspiration, the temperature which is actually felt depends to a great extent upon the dryness of the air, or, to put it more plainly, upon the reading of the wet-bulb thermometer. The evaporation of perspiration results in a cooling of the surface of the body which may be of very considerable amount; and this cooling is indicated by the wet-bulb thermometer. For this reason, the surface of the skin has been compared with a wet-bulb thermometer, and the temperatures indicated by the latter are taken as representing the temperatures which are actually experienced.[1]

that the quotient of body temperature minus air temperature, divided by skin temperature minus air temperature, is practically constant, and amounts to 1·4°. The equation is:

$$(37{\cdot}6 - t) \div (T - t) = 1{\cdot}4\,; \text{ hence, } T = 26{\cdot}5 + 0{\cdot}3t\,;$$

therefore, the skin temperature at 0° = 26·5°, at 20° = 26·5° + 6 = 32·5°, assuming that the air is calm.

The influence of the velocity of the wind in lowering the temperature of the skin was reckoned by Vincent, on the basis of his observations, as $1{\cdot}2° \times v$, and the influence of insolation as $0{\cdot}2°\ d$, where d represents the excess of temperature registered by the actinometer over that of the air (t). Thus he comes to the expression:

$$\text{Temperature of the skin} = 26{\cdot}5° + 0{\cdot}3°\ t + 0{\cdot}2°\ d - 1{\cdot}2°\ v.$$

These figures naturally represent only a first attempt to determine the temperature of an exposed surface of the skin by using the air temperature, the temperature in the sun, and the wind velocity. Nevertheless, they should be included in a text-book of climatology. The temperature in the sun is probably the most difficult to define. Vincent took the readings of a black-bulb thermometer in a vacuum enclosed by a glass bulb.

[1] N. P. Schierbeck has recently shown that the rate of evaporation is proportional to the square root of the wind velocity, as de Heen had previously assumed. If v denotes the volume of water vapour (at 0° C., and at a pressure of 760 mm.) formed in a unit of time; e the vapour tension in the air; e' the maximum vapour tension at the temperature of the evaporating water; k a constant: W the velocity of the wind, then (if T is the so-called absolute temperature, *i.e.*, $T_0 = 273$)

$$v = k(e' - e)\sqrt{W}(T \div T_0).$$

According to Trabert's view, this formula, which is based upon Dalton's assumption that v is proportional to the difference $e' - e$, probably best expresses the requirements of theory as well as the facts of observation. As $e' - e$ in the psychrometer formula is proportional to the difference between the dry and wet bulb readings, the rate of evaporation is likewise proportional to the latter. The difference between the temperatures of the dry and wet bulb thermometers is therefore also an approximate measure of the rate of evaporation. For this reason, the reading of the wet-bulb thermometer, or the difference between wet and dry bulb thermometers, attains added importance. A relative humidity of 40 per cent., which is often noted in heated rooms in winter, is easily borne indoors, whereas the same humidity out of doors would be disagreeable. If $W = 4$ meters

It is well known that very high temperatures can easily be endured when the air is dry; on the other hand, when the air is moist, the conditions are favourable for sunstroke. Sunstroke is by no means uncommon in the eastern United States in summer; but it is almost unknown in the dry West, *e.g.*, in Arizona and Southern Colorado, although the temperatures are much higher in the latter regions. The inhabitants of the eastern coast, says General Greely, are amazed at the temperatures of 45° to 50°, which occur in the West without interfering to any considerable extent with the ordinary occupations in city or country. In one of the hottest places in the world, namely, Death Valley, California, the observers of the Weather Bureau experienced five days in the summer of 1891, on which the maxima were 50°. The wet-bulb thermometer meanwhile indicated only 23° – 25°; so that the temperature actually felt by anyone who was in a favourable location, that is, protected against heating by radiation, was almost as low as that of an ordinary summer afternoon. On August 4 and 5, the maxima were 47·7° and 45·5°, but the dew-point was – 1° and – 2·8°; and the reading of the wet-bulb thermometer was 21° and 19·4°. It was therefore relatively cool.[1]

Harrington has called the temperatures indicated by the wet-bulb thermometer *sensible temperatures*, and he emphasises their importance in climatology.[2] He has drawn a chart of the July sensible temperatures for the whole of the United States, and another which shows the reduction of the temperature of the air by evaporation in the same month. On the latter chart the line of 2·8° encloses the eastern and southern United States, while in the interior of the West there is a considerable area which is enclosed by the line of 11° – 12°, that being the amount by which the air temperature is reduced by evaporation. The mean sensible July temperature in the eastern United States is 18·3° at Boston, to 24·4° at Savannah; while in the hot West, where the July isotherms are 30° – 34°, the July sensible

per second, a relative humidity of 64 per cent. out of doors corresponds to 40 per cent. in the calm air of a room. When the wind velocity is 13·7 meters per second, then 80 per cent. outdoors corresponds to 40 per cent. indoors ("Ueber die Geschwindigkeit der Verdampfung vom speziellen Gesichtspunkt der physiologischen Beziehungen." Oversight K. Dansk. Vid. Kab. Selsk. Forh., 1896, No. I.). See also W. Trabert: Neuere Beobachtungen ueber die Verdampfungsgeschwindigkeit. *M.Z.*, XIII., 1896, 261-263.

[1] M. W. Harrington: "Climate and Meteorology of Death Valley, California." *U.S. Dept. of Agriculture*, *Weather Bureau*, *Bulletin* No. I., 1892.

[2] M. W. Harrington: "Sensible Temperatures," *International Medical Magazine*, August, 1894.

temperatures are only 15·5° to 21°, the latter being the mean at Yuma. So far as the feeling of heat is concerned, the summer in the New England States is therefore hotter than that in the deserts of Arizona and southern California. Moore[1] has published a chart showing the average actual and sensible temperatures of the United States for the summer season, deduced from eight years' observations at 8 A.M. and 8 P.M., 75th meridian time. The great interior valleys and the plains east of the foothills of the Rocky Mountains have an average temperature ranging from 18·3° on the northern boundary to about 26·7° on the Gulf coast. On the other hand, the sensible temperatures west of the 105th meridian range from 10° to 18·3° and east of the 105th meridian they range from 12·8° to 23·9°. The line which marks the temperature of evaporation in New England and the Great Lakes, 15·6°, runs almost due north and south along the eastern foothills of the Rocky Mountains, and skirts southern New Mexico and Arizona. The 12·8° line passes almost due south from eastern Montana to south-eastern New Mexico, and thence northwesterly. The temperature of evaporation in all the territory above this line of 12·8°, embracing almost two-thirds of the arid region, is below 12·8° and almost one-third of the region is not over 10°. In the case of hot climates it would be advisable to include, among the climatic elements, the readings of the wet-bulb thermometer as a convenient index of the degree of heat which is actually felt by the human body.

[1] W. L. Moore: "Some Climatic Features of the Arid Regions," *U.S. Department of Agriculture, Weather Bureau*, 8vo, Washington, D.C., 1896.

CHAPTER II.

THE MOISTURE OF THE ATMOSPHERE: HUMIDITY, PRECIPITATION AND CLOUDINESS.

Absolute humidity. The psychrometer.—Next to the temperature, the climatic element of most importance is the measure of humidity, either as water vapour, or as water in the form of clouds, rain, snow, etc. The essential features of the humidity of any place are described when the quantity of water vapour in the air, and the amount and kind of precipitation are given.

The instrument which is used almost exclusively for the determination of atmospheric humidity is the psychrometer.[1] Observations of humidity by means of this instrument give us first of all the vapour pressure. This vapour pressure indicates the amount of water in the air, and is the value which must be included in all physical, and particularly in all meteorological, problems which are concerned with the humidity of the air. The term vapour pressure has, however, often led to misinterpretation. For instance, the vapour pressure has been deducted from the air pressure as measured by a barometer, and the resulting figures have been supposed to express the "weight of the dry air." This mistake is still occasionally made. It has also been supposed that the vapour pressure is a measure of the total amount of water vapour in the atmosphere overhead, just as the weight of the air column above the place of observation is indicated by the height of the barometric column. These views are erroneous, as the author has

[1] See also L. E. Jewell: "The Determination of the Relative Quantities of Aqueous Vapor in the Atmosphere by means of the Absorption Lines of the Spectrum," *U.S. Department of Agriculture, Weather Bureau, Bulletin* No. 16, 8vo, Washington, D.C., 1896, pp. 12. fig. 1. Also *Astro-Physical Journal*, IV., 1896, 324-342.

already shown at length,[1] and hold good only in the case of small enclosed spaces in which it may be assumed that the water vapour has, by diffusion, already become uniformly distributed, and where no condensation of this vapour takes place. This condition can never occur in the free air. According to the older view, of which Dove was the principal champion, it must be held that, with a mean vapour pressure of 11 mm. at Vienna, in August, the water vapour in the air under these conditions would, if wholly condensed, give a layer of water $11 \times 13{\cdot}6 = 149{\cdot}6$ mms. deep.[2] It has, however, been shown by the author that the actual amount of water in the atmosphere at this time is in round numbers only one-fifth, or, to be more exact, 0·22, of the amount as determined by the foregoing calculation.

There is some temptation to replace the vapour pressure, which is so apt to lead to error, by the weight of the water vapour in unit-volume, preferably in a cubic meter (or a cubic foot). This would certainly be the most intelligible, as well as a very practical, expression for the vapour contents of the air. In fact, English climatic tables very commonly give the weight of water vapour, expressed as so many grains in a cubic foot. Nevertheless the author is not disposed to recommend this method, and the reasons against its adoption will appear.[3]

A most fortunate coincidence, which occurs when the metric system is used, may, however, be noted here. In the case of the temperatures that usually occur out of doors, the vapour pressure in millimeters, and the weight of water vapour in grams per cubic meter, are expressed by almost the same figures.[4]

[1] J. Hann: "Die Abnahme des Wasserdampfgehaltes der Atmosphäre mit zunehmender Höhe," *Z.f.M.*, IX., 1874, 193-200. Lamont, at a much earlier date, unsuccessfully opposed these ideas when they were upheld by Dove.

[2] Vapour pressure, like atmospheric pressure, is measured by the height of a mercury column of equal weight. Therefore, in order to obtain the depth of water, it is necessary to multiply by the specific gravity of mercury.

[3] In some meteorological calculations and investigations, the weight of water vapour in unit weight (*e.g.*, the kilogram) is of importance. Von Bezold calls this weight the *specific humidity*.

[4] When e is the vapour pressure, p the weight of the water vapour in grams per cubic meter, and t the temperature of the air, the relation is as follows:

$$p = 1{\cdot}06e \div (1 + 0{\cdot}0037t).$$

Below 16° the denominator is smaller than the numerator, 1·06, and the weight of the water vapour is therefore somewhat greater than e. Above 16° the relation is reversed. Tables for obtaining p easily can be found in Jelinek's

The vapour pressure therefore also gives an expression, which is of sufficient accuracy for most purposes, for the weight of the water vapour in a cubic meter of air.

Vapour pressure, or the weight of the water vapour, is also known as the *absolute humidity.*

Relative humidity. "Saturation-deficit."—In contrast with this is the relative humidity, *i.e.*, the degree of saturation of the air with water vapour, or the ratio between the observed vapour pressure and the vapour pressure at saturation under existing conditions of temperature.

It has recently been urged by several investigators that atmospheric humidity can, for most practical purposes, be much better expressed by the difference between the observed vapour pressure and the maximum vapour pressure possible at the temperature then prevailing. This difference is known as the *saturation-deficit.*[1]

Dew-point.—In addition to the data which have already been mentioned as expressions for the humidity of the atmosphere, namely, the absolute and the relative humidity, and the *saturation-deficit*, there is one more quantity[2] which may also be regarded as a measure, although

Anleitung zu meteorologischen Beobachtungen, Part II., 4th Edition (Leipzig, 1895), p. 23. The following condensed table will serve for most purposes :—

Temperature .	−10°	0°	5°	10°	15°	20°	25°	30°
Factor, . .	1·100	1·060	1·040	1·022	1·005	0·987	0·971	0·955

The observed vapour pressure, e, is to be multiplied by these factors in order to give p.

[1] The discussion by H. Meyer ("Untersuchungen ueber das Sättigungsdefizit," *M.Z.*, IV., 1887, 113-124), is worthy of special attention from a practical climatological standpoint. The writers who have recommended the use of this new expression are Fluegge (*Lehrbuch der hygienischen Untersuchungsmethoden*, Leipzig, 1881); T. Denecke (*Zeitschrift für Hygiene*, Vol. I.), and Meyer, (*M.Z.*, II., 1885, 153-162, and IV., 1887, 113-124). If E denotes the vapour pressure at saturation, e the observed vapour pressure, the relative humidity is the quotient of $e \div E$, and the *saturation-deficit*, the difference $E - e$. The *saturation-deficit* is therefore equal to (1 − relative humidity) E. Thus it is evident that with the same relative humidity the *saturation-deficit* increases with the temperature. This increase is, furthermore, very rapid, because E increases very rapidly with the temperature. Denecke and Meyer, in their writings above referred to, have given tables by means of which the *saturation-deficit* can be derived from the air temperature and from the relative humidity by a still simpler method.

[2] The physicist Jamin has, it is true, made the suggestion, which is an impractical one, that climatic tables should, in place of the relative humidity, include the quotient $e \div (b - e)$, in which b is the height of the barometer, and e the vapour pressure. This is the ratio between the weight of the water vapour and the weight of the air, the volumes being equal. The ratio is always a very small quantity,

an indirect one, of atmospheric humidity. This is the dew-point, *i.e.*, the temperature down to which the air must be cooled in order to cause a condensation of the vapour in it. The condensation hygrometers of Daniell, Regnault, Alluard, Crova and others, indicate nothing but the temperature of the dew point. The vapour pressure and the remaining data must then be determined. The dew point alone gives no direct indication as to the amount of vapour in the atmosphere, but the difference between the existing air temperature and the temperature of the dew point furnishes an indication of the degree to which the air is saturated with water vapour. The greater this difference, the drier is the air. The readings of the wet-bulb thermometer are useful additions to the climatic tables of warm countries, because these readings are an important index of the temperatures which are actually felt, as has already been noted.

Absolute and relative humidity compared.—The absolute humidity, expressed either as vapour pressure, or as the weight of the water vapour in a unit volume, always suffices for the physicist and the meteorologist; but as an expression of the climatic humidity, or as an indication of the effect of this humidity upon the organism, the absolute humidity alone is of no value. Air can be very "dry" and yet contain more water vapour than "very damp" air, provided the temperatures differ considerably in the two cases. Rohlfs found the mean vapour pressure in the Kufra Oasis, in the heart of the Libyan Desert, 8·3 mm. in the second half of August, and 11·1 mm. in September, which corresponds to the absolute humidity of Vienna and Oxford (England) in summer. The lowest figures for the vapour pressure were 4·5—5·5 mm., and these correspond to the average amount of vapour in the damp winter air of England. On August 14, at 3 P.M., the temperature of the air at Hauari (Kufra Oasis) was 38·9°; the wet-bulb thermometer indicated 18·9°; the vapour pressure was 4·5 mm.; the relative humidity, 9 per cent.;

and the quotient must be carried out to at least the fourth decimal place. The average vapour contents of the air at Vienna would, for example, be 0·0094. Furthermore, the true meaning of the *richesse hygrométrique*, as Jamin calls this quotient, is not always perfectly clear, and Jamin himself has misused the term. This quotient can be but seldom employed in meteorological calculations, and can never replace the relative humidity, or the *saturation-deficit*. Notwithstanding the laborious figuring connected with its determination, the French have begun to include this useless *richesse hygrométrique* in their meteorological tables. (Cf. the author's criticism in *Z.f.M.*, XIX., 1884, 408-410, and XX., 1885, 269-272.)

the saturation-deficit, 47 mm.; the dew-point, $-0{\cdot}2°$; the wind, light, from E.N.E. That is typical desert dryness, in which finger nails split and the skin peels off. Yet the absolute amount of water vapour in the air was still equal to that of the damp winter air of Western Europe, although, in the latter case, the relative humidity is 80-90 per cent.

Since air breathed into the body is warmed by the lungs to the temperature of the body, and is breathed out again filled with water vapour, it has been maintained by some that absolute humidity plays a more important part as a climatic factor than relative humidity. When the air is perfectly dry, the body loses more than 430 grams of water through the process of respiration every day. This corresponds to saturation at 37° and 10 cubic meters of air per day. Yet air of this degree of dryness is breathed out of doors all winter long in Eastern Siberia and in Arctic districts; for even in saturated air the vapour pressure at $-30°$, $-40°$, and $-50°$ is only 0·4, 0·1, and 0·04 mm. respectively. The air is almost absolutely dry, and yet no complaints are heard about the dryness of the air, and there is no mention of its effects.

An able physicist and naturalist, Ermann, travelled during the winter through all Western and Eastern Siberia, as far as Ochotsk, experiencing temperatures of $-40°$ to $-50°$ and below. He describes the effects of the cold, but makes no mention of the effects of the absolute dryness of the air. The same thing is true of von Middendorff and others. Payer alone speaks of "torturing thirst," but it must be remembered in this connection that Payer and his companions themselves dragged the sledges with their supplies, etc., and that their thirst was therefore the result of their great physical exertions.[1] At the same point in his narrative, Payer does, however, speak of the feeling of penetrating dampness, which is all the more noticeable when the temperature is very low. This statement of Payer's had reference to Franz Joseph Land, where the relative humidity was very high, owing to the proximity of bodies of open water. This proves clearly enough that even a considerable amount of water may be given off to the air in respiration while the loss is hardly perceptible to the body. It would, however, be interesting in this connection to examine medical reports from places in Siberia, where the mean winter temperatures are between $-20°$ and $-50°$.

The case appears to be different when the air which is breathed in already contains so much water vapour that the loss of water from the lungs is checked. This seems to affect the body much more acutely, but there is nowhere, under natural conditions in the atmosphere, relatively dry air which contains so large an amount of water vapour. When air with the same absolute humidity is nearly saturated (that is, when the relative humidity is also high), then evaporation from the surface of the skin is also checked, and the air seems extremely muggy. It is impossible, under these conditions, to tell how much of the effect is due to the reduced loss of water from the lungs, and how much to

[1] Every one has undoubtedly noticed, in his own case, that the feeling of thirst out of doors is not nearly so great in winter, with temperatures of $-10°$, or $-20°$, as it is in summer with a temperature of 30°.

the reduced loss from the skin. Dr. Fleischer believes that sultry air has its dew-point at 19°.[1]

For purely climatological purposes the relative humidity is, unquestionably, the most convenient expression for the amount of water vapour in the air. When we describe the air as being damp, or dry, we are usually speaking quite unconsciously of the relative humidity. The air is moist in our climate in winter, notwithstanding the small amount of water vapour which it then contains; while the air is dry in summer, although it then contains two or three times as much vapour as in winter. The relative humidity, next to the temperature, determines the need which is felt by organisms for water, and also controls evaporation.[2]

The relative humidity is, furthermore, by no means an expression which is used only in computations. It is a perfectly definite climatic factor, as can be seen from the fact that it is directly indicated by organic substances. All organic substances are more or less hygroscopic, and their condition, so far as it depends upon the humidity of the air, is determined by the relative, and not by the absolute, humidity.[3] Thus it happens that organic substances, such as membranes or hairs, furnish us with excellent means for the direct measurement of the relative humidity of the air. All other measurements of humidity are indirect, and involve a somewhat difficult calculation, the results of which are in certain respects less accurate than those obtained by means of the hair hygrometer. The readings of the psychrometer below freezing are a case in point. The relative humidity is therefore the most natural expression for the humidity of the air as a climatic factor, for it reacts directly upon organic substances.[4]

[1] *Gesunde Luft*, Göttingen. See also A. Lancaster: "De la Manière d'utiliser les Observations hygrométriques," *Rapport lu au Ve Congrès International d'Hydrologie, de Climatologie et de Geologie Medicales de Liège*, 1898, and the results of observations by Rubner and von Lewaschew (*Archiv für Hygiene*, XXIX., 1897, 1), [*M.Z.*, XV., 1898, 148-149.]

[2] E. G. Ravenstein has recently considered the geographical distribution of relative humidity (*Rept. Brit. Assn. Adv. Sci.*, 1900, 817-818).

[3] In this connection see A. J. Henry: "Report on the Relative Humidity of Southern New England and Other Localities," *U.S Dept. of Agriculture, Weather Bureau, Bulletin* No. 19, 8vo. Washington, D.C., 1896, pp. 23, pls. IV.

[4] Sresnewsky has recently proposed a mathematical theory of the hair hygrometer. He shows that the tension of the fluid in the pores of the hair is directly proportional to the natural logarithms of the relative humidities. He gives a formula which expresses the relation between the relative humidity and the changes in the length of the hair (Dorpat, 1895).

As the relative humidity usually varies greatly during the day, it is necessary, in climatic tables, to give not only the mean for the day, but also the means for the different observation hours by themselves, that is, for morning, afternoon, and evening. (See Table, page 81, columns 1-5.) It is also important to include the minima of the humidity. From a hygienic point of view it would further be of great interest to have the variability of the relative humidity from day to day, as shown by the daily means, or, better still, by the afternoon means, also included in descriptions of the climates of health resorts. The absolute humidity has so slight a daily range that the monthly means obtained from the three observation hours suffice.

Relative humidity and saturation-deficit compared.—Objection has lately been made to the use of the relative humidity, on the ground that it does not properly express the influence of atmospheric humidity upon the body, and it has been proposed that the saturation-deficit should be substituted. It is perfectly true that the rapidity of evaporation is much more nearly proportional to the saturation-deficit than to the relative humidity, while it has been objected to the use of the relative humidity that the same percentage of relative humidity has very different meanings when the temperatures differ. Thus, for example, a humidity of 30°/₀, when the temperature is 25°, is not the same as a humidity of 30°/₀ when the temperature is – 10°; and the effects upon the body are also different in the two cases. The saturation-deficit, on the other hand, is independent of the temperature. It is further urged that, in order to have a given percentage of relative humidity really mean what it should, it is always necessary also to take into account the temperature prevailing at the time, which is not the case in relation to the saturation-deficit.

It may be admitted that the introduction of the saturation-deficit into climatic tables would not be altogether undesirable, in view of the fact that its practical importance might thereby be determined. It might then be possible to decide the question whether the saturation-deficit really should be preferred to the relative humidity. The psychrometer tables in ordinary use give directly only the vapour pressure and the relative humidity. It is, however, a simple matter to obtain from these same tables the vapour pressure when the air is saturated at the temperature indicated by the dry-bulb thermometer, (*i.e.*, E), and to subtract from this the prevailing vapour pressure, which gives $E - e$, the saturation-deficit. Furthermore, the means of the air temperature and of the vapour pressure, and even of the relative

humidity, serve as a basis for determining the saturation-deficit, and, although the values of the latter when thus obtained are not absolutely accurate, they are sufficiently so for practical purposes. Some writers have even gone so far as to say that the relative humidity is of no practical importance, and have proposed to omit it altogether from climatic tables. It therefore seems necessary to enter a protest against this hasty judgment, and to point out that there are very excellent reasons against giving the values of the saturation-deficit alone, and that in most cases the relative humidity gives a better idea of the "climatic humidity" than does the saturation-deficit. In the first place, the saturation-deficit does not indicate the "climatic humidity" when the air temperature is not given, and hence it does not follow that therein lies the great advantage of the saturation-deficit over the relative humidity. In the case of the former element, as in that of the latter, the same reading has values which vary greatly with different temperatures. This can readily be proved. Our bright winter days, with sharp frost and brisk northerly or easterly[1] winds, are properly to be called dry from the point of view of their physiological effects. The air then has an enlivening, stimulating quality; in short, it has the characteristic effects of dryness. It is hardly necessary to point out that this is not alone due to the low temperature, because, when the temperature is the same, but the air is damp, the effects are quite different. Yet the saturation-deficit is then much smaller than it is in summer, when the temperature is high and the atmosphere is very muggy and oppressive. Every attentive observer is aware of the drying qualities and the peculiarly injurious influences of the dry air on some of our spring days. Yet the saturation-deficit may then be smaller than that of the hot, damp air of the tropics. In the latter, iron rusts quickly; mould soon forms on leather, and the Northerner is exposed to all the injurious effects of a warm atmosphere nearly saturated with water vapour. In such cases the relative humidity is a much better indication of these effects than is the saturation-deficit.

In the Arctic regions the air may be very dry in winter, although the saturation-deficit must always remain very small on account of the low temperature; for, at a temperature of $-20°$, the vapour pressure at saturation is only 0·9 mm., and at $-30°$ it is only 0·4 mm. Notwithstanding this fact, we read concerning eastern Greenland, where the winter temperature is $-20°$: "The air was very dry (in winter). The breath did not appear as fog even in the most severe cold, although this phenomenon usually occurs at temperatures above freezing in our damp climate" (*Second German North Polar Expedition*, II., 533).

On the other hand, Payer complains of penetrating cold in Franz Joseph

[1] In *Europe*.

Land, and adds, "The feeling of dampness was less uncomfortable during our sledge journeys in Greenland because of the lower percentage of relative humidity in the latter case." Eastern Greenland has strong prevailing winds from N. and N.W., which are relatively dry, while Franz Joseph Land has relatively damp winds from the ocean. The saturation-deficit was undoubtedly smaller in Franz Joseph Land, on account of the lower temperature [winter mean, – 28°], but the humidity was nevertheless greater. As Payer says, "The breath comes steaming out of the mouth, and surrounds the traveller with a cloud of fine ice-needles.'

Von Middendorff gives a very striking description of the dryness of the winter air in Eastern Siberia. He also says that it would probably be impossible for man, with his nomadic habits of life there, to endure the extreme cold of Eastern Siberia unless he were helped by the dryness of the air. Fur clothing which has become damp during the day through the evaporation from the body, is laid inside out on the snow during the night, and in the morning it is completely dry.[1] This happens with a saturation-deficit of 0·0 mm. !

In Batavia the conditions on January 23, 1882, at 2 P.M., were as follows: Temperature, 29·8°; saturation-deficit, 10·9 mm.; relative humidity, 65 per cent. These conditions indicate an atmosphere which is damp and oppressive, at least for Europeans and Americans. The author looked through the Batavian observations, and found that during the excessively damp rainy season, which occurs there from December to February, there is often a saturation-deficit of 6 – 8 mm. when the air is very muggy, *i.e.*, when the temperature and the relative humidity are high. January, which is both hot and damp, and which has a rainfall of 400 mm., has a relative humidity of 88 per cent., and a saturation-deficit of 3·0 mm., with a mean temperature of 25·3°; while April, in Vienna, which has a dry atmosphere, with a mean temperature of 9·7°, and a relative humidity of 66 per cent., has almost the same saturation-deficit, namely, 3·1 mm. The humidity in Batavia in the middle of the rainy season, when expressed by the saturation-deficit, is the same as it is in Vienna in spring. At this time, according to Junghuhn, the air is so damp in Batavia that people are at a loss to know how to keep their things from being ruined by the dampness, even indoors.[2] It is thus very clear that the saturation-deficit cannot express the conditions of a hot and damp tropical climate, nor the mugginess of our summer days.

G. Schott also comes to the conclusion that the saturation-deficit often gives a very false impression of the atmospheric humidity.[3] Of the examples which he gives in support of his view we shall cite

[1] *Reise im äussersten Norden und Osten Sibiriens*, IV., 393, *et sqq.*

[2] The reader is further referred to the eloquent description of the sufferings of Europeans during the rainy season in Senegambia, especially in St. Louis, given by Surgeon-General of the Navy, D. A. Borius (*Les Maladies du Sénégal*, Paris, 1882, 124, *et sqq.*). The mean temperature is then 27·6°, and hardly goes above 31°; the mean saturation-deficit is 5·5 mm., as in central Europe in May and June; but the relative humidity is nearly 80 per cent.

[3] "Wissenschaftliche Ergebnisse einer Forschungsreise zur See." *Pet. Mitt.*, Ergänzungsheft 109, 1893, 121-124.

but one. "In the belt of calms of the Indian Ocean the saturation-deficit was 5·0 mm., with a relative humidity of 81 per cent., and a vapour-pressure of 21·5 mm., while the corresponding figures in the trade belt of the Indian Ocean were 4·5 mm., 79 per cent. and 18·6 mm. Judging by the saturation-deficit, therefore, the air in the trades would have been more moist than that in the belt of calms, which is, of course, far from being the case." The reason that the humidity, as expressed by the saturation-deficit, gives such a false impression, is that the temperature in the belt of calms was 27·0°, but in the trades it was only a little above 25°. This slight difference of temperature is enough to cause air which is really drier to appear moister when the humidity is indicated by the saturation-deficit.

These special examples may suffice to show that the saturation-deficit does not give any accurate idea of the climatic humidity, unless the corresponding air temperatures are also taken into consideration. In fact, it is inferior to the relative humidity in this respect. It would, therefore, by no means be desirable to substitute the former for the latter. The use of the saturation-deficit *in addition to* the relative humidity can be recommended only with a view to studying its usefulness in hygiene and therapeutics. The saturation-deficit has the further disadvantage as a climatic factor, that it does not represent the humidity conditions of the air so directly as does the relative humidity. The conditions associated with a relative humidity of 50-80 per cent. can at once be clearly brought before our minds if we know the time and the place that we are dealing with, *i.e.*, if we know in a general way what the temperature is. This is not the case with a saturation-deficit of, say, 2 or 8 mm. Here the temperature must be accurately known if we are to be able to determine whether the air is climatically damp or dry; whether it is muggy and oppressive, or stimulating.

Physiological effects of varying degrees of relative humidity.—According to Pettenkofer and Voit, an adult man eliminates about 900 grams of water from his skin and lungs daily. Of this amount 0·6, or 540 grams come from the skin alone; and changes in relative humidity of only 1 per cent. cause perceptible changes in the amount of evaporation from the skin. If evaporation from skin and lungs is diminished, the amount of urine is increased, as, in many cases, are also the secretions from the intestines. For this reason sudden changes in the humidity have very injurious effects upon a body which is not in health. Such changes make themselves felt chiefly in the sudden increase or decrease of the pressure of the blood. For this reason, as has already been noted, the determination of the variability of the relative humidity

would be a necessary element in a complete description of a climate with reference to its relation to health. The less-diluted blood of dry climates operates as a stimulant upon, and increases the functions of, the nervous system. The consequences are excitement and sleeplessness. This effect, which manifests itself in a certain restlessness, is also seen in the case of persons who are in perfect health soon after they come into a dry climate, or a mountain climate. Mountain climates, even when they have a higher relative humidity, can in this respect be placed on a par with the dry climates of places at a lower altitude, because the diminished pressure at high altitudes increases evaporation.

Damp air, and increased pressure, have the following physiological effects:—nervous depression; quiet sleep; increased elimination of carbon dioxide; slower circulation of the blood. Dry air, and decreased pressure, on the other hand, have these effects:—Nervous excitement; sleeplessness; quickened pulse; a drier skin, and a decreased temperature.[1]

The physiological effects of changes in temperature vary according to the relative dampness or dryness of the air. When the relative humidity is high, a slight cooling is very noticeable and may also be injurious, while the reverse is the case when the air is dry.[2] The inhabitants of deserts and of dry regions in general endure, without the slightest discomfort, great changes of temperature which would be very harmful in damper climates. As regards their physiological effects, therefore, changes of temperature are not alike in different climates.

Travellers through the Sahara, especially Nachtigal, have given us very striking examples of the effects of the relative humidity upon the physical and mental characteristics of man, in the difference which exists between the inhabitants of the desert and those of the moist Soudan. In this case there is, to be sure, also a considerable difference between the absolute humidity in the desert and in the Soudan; but other examples indicate clearly enough that the relative humidity is the chief factor in the matter.

Rainfall data, including rain, snow, hail, dew and frost.—(See Table, page 81, columns 6 and 7.) The amount of precipitation is in some respects the climatic element of most importance after temperature. The rainfall determines the productiveness of a country. Temperature and rainfall together are one of the most important natural resources of a country. Wills gives the following table to

[1] Thomas: "Beiträge zur Allgem. Klimatologie." Erlangen, 1872.

[2] G. von Liebig: "Ventilation und Erwärmung in pneumatischen Kammern." Munich, 1869.

show the increase in the productive capacity of the grazing districts of Australia and Argentina with the increase in rainfall.

RELATION BETWEEN ANNUAL RAINFALL AND NUMBER OF SHEEP PER SQUARE MILE IN AUSTRALIA AND ARGENTINA.

District.	Rainfall.	Sheep per sq. mile.	Increased Capacity for every Inch of Rain, etc., added.
South Australia, . .	Say, first 8-10 inches,	8-9 ?	About 1 sheep per mile.
New South Wales (1), .	9+4 ,,	96	,, 22 ,, ,,
,, ,, (2), .	9+4+7 ,,	640	,, 70 ,, ,,
Buenos Ayres, . . .	9+4+7+14 ,,	2630	,, 140 ,, ,,

It appears further that in the wheat districts of southern Australia during the six winter months the difference in bushels per acre of average yield is almost exactly the same as the difference in inches of rain. The seven best years yielded 12·4 bushels with 18·5 inches of rain; the five next best years gave 10·0 bushels with 15·4 inches; and six bad years, which yielded 6·6 bushels, had 13·5 inches of rain. The average difference between the number of bushels and inches of rain is constant, and is just about 6. "Land without rain is worth nothing; and land in an Australian climate, with less than 10 inches a year, is worth next to nothing. Rain-water, without land, if the water can be stored in a reservoir and sent along a canal, is worth a great deal."[1]

Rawson has worked out a simple formula in the case of Barbadoes, by means of which the amount of sugar to be exported the next year can be calculated with great accuracy from the rainfall of the present year. This calculation is accurate within 6 per cent. in most cases.[2]

Similar calculations have been made for Jamaica. Fifty-six inches of rain give 1441 casks per acre; 76 inches give 1559, or about one-tenth more; which means an increase in the value of the sugar crop alone of £100,000.[3]

A study of the relation between the amount and distribution of the rainfall and the wheat crop in California has recently been made by

[1] J. T. Wills: "Rainfall in Australia," *Scot. Geogr. Mag.*, Vol. III., 1887, 161-173.

[2] "Report upon the Rainfall of Barbadoes, and upon its Influence on the Sugar Crops," 1847-1871. With two Supplements, 1873-1874, Barbadoes, 1874 (*Z.f.M.*, IX., 1874, 318-320).

[3] *Nature*, Vol. XXXI., 1884-85, 538.

Fraser,[1] and Clayton has discussed the influence of rainfall on commerce and politics in the United States.[2]

For climatological purposes, precipitation is recorded by means of the following data:

(*a*) The monthly and annual amounts of rainfall. This means the total depth of water resulting from the various kinds of precipitation already enumerated. Snow is melted and is then recorded as so many millimeters of rain.[3]

(*b*) The maximum precipitation per day, or per hour, or in any shorter period.[4]

(*c*) The number of rainy days, *i.e.*, of days having 0·1 mm. or more of precipitation. Days on which dew is deposited may therefore very seldom be included, at least in temperate zone climates. The number of days with a precipitation of 1 mm. and over is also important, because these data facilitate comparisons. The recorded number of days with small rainfalls (below 1 mm.) varies greatly with the care of the observers, and with the kind, size, and exposure of the rain-gauge. Therefore, it is very desirable, in order to have accurate comparisons, that the number of days with rainfalls of 1 mm. or more should be recorded.

The number of rainy days is a climatic element which should always be recorded in addition to the amounts of rainfall, because the number of days upon which the given precipitation has fallen is of the greatest importance to vegetation. Although the amount of precipitation may be very considerable, there may be great drought if the rain happened

[1] W. H. Fraser: "Rainfall and Wheat in California," *Overland Monthly*, XXIII., 1899, 521-553.

[2] *Pop. Sci. Mo.*, Vol. LX., 1901, 158-165.

[3] It is amply sufficient if the depths of rainfall are recorded to whole millimeters. See J. Hann: "Ueber die Reduktion kürzerer Reihen von Niederschlagsmessungen auf die längere Reihe einer Nachbarstation," *M.Z.*, XV., 1898, 121-133.

[4] See "Tables of Excessive Precipitation" in *Report of the Chief of the Weather Bureau* for 1895-96, 4to, Washington, D.C., 1896, pp. 247-266. These tables show, for a number of stations of the Weather Bureau, the maximum amounts of excessive precipitation for periods of five and ten minutes; one hour, and twenty-four hours or more, and also the accumulated amounts of excessive precipitation for each five minutes during some of the heaviest rainfalls at stations furnished with self-registering gauges. See also Symons on maximum rainfalls in twenty-four hours, in *British Rainfall*, 1895, 40 (*M.Z.*, XVI., 1899, 26-27), and R. H. Scott: "On the Heavy Falls of Rain recorded at the Seven Observatories connected with the Meteorological Office, 1871-1898," *Quart. Jour. Roy. Met. Soc.*, XXV., 1899, 317-323.

to fall on one day, or on a few days only, while the remaining days were rainless, with prevailingly high temperatures.

Soil-moisture and character of the precipitation.—In his investigations concerning the relation between the moisture of the soil and precipitation, Wollny emphasises the importance of *heavy* rainfalls, as contrasted with *more frequent* rainfalls, which are of little benefit because they evaporate too quickly. He believes that a daily rainfall of 2 mm. in summer is of less benefit than the corresponding amount of 180 mm. coming in ten or twelve showers.[1] Wollny seems, however, to have overlooked one point, viz., that the small rainfalls which occur frequently are almost always widely distributed. They are, therefore, associated with a prevailingly high atmospheric humidity and continued cloudiness, and the ground is thus protected from evaporation. The case is different with short but heavy downpours. Most of the water which has fallen runs off on the surface, and as these showers are usually only local, the air almost immediately becomes dry again, and the sun quickly parches the ground. It is well known that the lower layers of the soil do not become wet after heavy rains, but that the surface is beaten down by the rain like a threshing-floor, and then dries quickly. On the other hand, the small rainfalls which come frequently soften the ground, and thus gradually furnish some water to the lower strata, or at least preserve the supply which is already there. The case is, of course, different when this subject is studied experimentally. Frequent and gentle artificial watering of the soil cannot bring about the prevailingly high atmospheric humidity and the heavy clouds which are important accompaniments of frequent light rainfalls. It is perfectly natural that an occasional watering should be the most effective in experiments of this kind. Furthermore, the effect of precipitation upon a sloping surface, which is the usual condition in nature, is very different from that upon one which is level.

Duration of individual rainfalls.—Woeikof rightly emphasises the importance of the *duration of rainfalls*,[2] but it is unfortunately difficult to obtain comparable data concerning this element of climate. Köppen has sought to suggest a method of removing this difficulty.[3]

The frequency of rainy days and the duration of the rainfalls need not necessarily correspond at all, although they certainly do correspond, according to the records, in the case of Paris and of Perpignan. Paris has 169·5 rainy days, and 654 hours of rain; Perpignan has 84·3 rainy days, and 312 hours of rain. The number of rainy days and of hours of rainfall at Perpignan are thus about one-half of the corresponding

[1] *Forschungen auf dem Gebiete der Agrikulturphysik*, XIV., 1891, 143.

[2] *Die Klimate der Erde*, I. 32.

[3] W. Köppen: "Zur Charakteristik der Regen in Nordwesteuropa und Nordamerika," *M.Z.*, II., 1885, 10-24. The article also contains interesting climatic tables. Cf., further, H. Meyer: "Die Niederschlagsverhältnisse Göttingens," *M.Z.*, IV., 1887, 415; "Die Niederschlagsverhältnisse von Deutschland insbesondere von Norddeutschland in den Jahren 1876 bis 1885," *Archiv der deutschen Seewarte*, XI. (1888), No. 6, 1889, pp. 45.

numbers at Paris. The number of hours with rain per rainy day is the same in both places, namely, 3·8. In general, it may be assumed that a rainy day in the Mediterranean climates means a smaller number of hours of rain than in central Europe.

Mean frequency of days with precipitation of a certain amount.—Great interest also attaches to the *mean frequency of days with precipitation of a certain amount*, as, for instance, 5, 10, 20, 30, 50 mm., and so on. It has been suggested that in countries which use the metric system, the special value of 25 mm. (= 1 inch) be used, as well as the even numbers, 10, 20, 30, 40 mm. etc., because comparisons with the data given in English tables could then be made. It is still better to enumerate the days with given amounts of rainfall according to groups, as, for instance, those with 1-5, 5, 1-10 mm., etc. In this way Hoppe has conveniently distinguished between days with light (up to 1 mm.), moderate (1·1-5 mm.), heavy (5·1-10 mm.), and very heavy (over 10 mm.) rain. Such summaries, and the means based upon them, certainly serve to give a very accurate idea of the rainfall of any district, and are of practical importance.[1]

Hinrichs[2] gives a good illustration of the value of such data for the case of Iowa. The rainfall of the year 1890 in that State was much more favourable to agriculture than that of the year 1899, although the annual amount in the two cases was about the same. The character of the rainfall of the two years, as analysed in the following table, explains this difference.

ANALYSIS OF RAINFALL AT IOWA CITY
IN 1889 AND 1890.

	1889.	1890.
Total rainfall, . . .	724 mm.	687 mm.
Washing and flooding rains, .	330 ,,	149 ,,
Insignificant rains, . .	36 ,,	29 ,,
Total utilisable rains, . .	358 ,,	509 ,,

Probability of rainy days.—By dividing the average number of rainy days in a month, or in a shorter period, by the total number of days in that period, we obtain an expression for the *probability of rain* during the interval in question. Thus, Vienna has, on the average,

[1] Cf. Meyer: *Anleitung*, p. 132, *et sqq.*

[2] "Rainfall Laws deduced from Twenty Years of Observation," *U.S. Weather Bureau*, 8vo, Washington, D.C., 1893, 15.

13·3 rainy days in July; the probability of rain is therefore ·43. In September there are only 8·3 rainy days; the probability of rain is therefore ·28. At Lesina, the probability of rain in summer is only ·10. While we can, therefore, count on four rainy days out of every ten at Vienna in July, at Lesina there is but one rainy day in every ten in summer.

Data concerning the probability of rain add very much to the completeness of records which simply give the amounts of rainfall, and are very important from a botanical standpoint, as well as in all questions relating to agriculture.[1]

Rain intensity.—If the total monthly rainfall is divided by the number of rainy days, an expression is obtained for the intensity of the rainfalls. Still better values for the intensity of the rainfalls would be obtained by dividing the total monthly rainfall by the number of hours with rain. Thus, for example, Paris has a mean annual rainfall of 574 mm.; a rain intensity per day of 3·4 mm. (4·4 mm. in August), and per hour of 0·9 mm. (1·6 mm. in August). Perpignan, on the other hand, with an annual rainfall of 598 mm., has a rain intensity per day of 7·1 mm. (12·1 mm. in December), and per hour of 1·9 mm. (3·0 mm. in August). The intensity of the rainfalls is twice as great in Perpignan as in Paris.

Additional precipitation data, which would be of importance in detailed studies of the climatic elements of any region, are the mean frequency of long dry and rainy periods, as well as of the longest of such periods. This involves the examination of meteorological records and the determination of the numbers of successive days with or without precipitation. The results should be classified according to seasons. Harrington has constructed charts showing the greatest number of consecutive days with and without rain and snow in the United States,[2] and F. H. Brandenburg has recently prepared[3] a list of all dry periods of 20 days or longer, arranged by seasons, which occurred at Denver, Colorado, between November, 1871, and December,

[1] For the United States, see "Charts showing the Monthly Probability of Rainy Days in the United States," *U.S. Signal Service*, 1891, and "Charts showing Annual and Greatest and Least Probability of Rain," in M. W. Harrington: "Rainfall and Snow of the United States," *U.S. Weather Bureau, Bulletin C.*, 1894; also, G. Hinrichs: "Rainfall Laws deduced from Twenty Years of Observation," *U.S. Weather Bureau*, 8vo, Washington, D.C., 1893, pp. 94 (reviewed in *Am. Met. Journ.*, X., 1893-94, 522-526).

[2] *Rainfall and Snow of the United States*, Chart XXII., 9 and 10.

[3] *Climate and Crops*, Colorado Section, November, 1899.

1899, inclusive. A dry spell is considered to be one in which not more than 0·01 inch of rain falls. Thirty-five such spells, of from 20 to 46 days' duration, are enumerated during the months from August to December; 21 cases, of from 20 to 58 days each, during the months from November to February; 10 cases, of from 20 to 28 days each, during the spring months, from February to May; and 5 cases, of from 24 to 50 days each, during the summer, from May to September.

The *number of days with snow* should be noted, besides the number of rainy days, and also the *depth of snow*. Further, the *duration of the snow covering* and the *average dates of first and last snowfall* are desirable climatic data. The *number of days with hail* and *the number of days with thunderstorms* should also be mentioned here. It has been agreed that the number of days with thunderstorms, and not the number of thunderstorms, shall be recorded. Although these latter statistics are not an important climatic element, they help to make the picture of a climate more complete.

Cloudiness.—The portion of the sky covered by clouds is an important datum. This is expressed in tenths or hundredths of the whole sky, so that a cloudiness of 63, or 6·3, for instance, indicates that 63 per cent. of the whole sky was covered by clouds (see Table, page 81, column 8). A simple statement as to the number of clear, cloudy and overcast days only, which is still quite a common method of indicating cloudiness, does not seem to be a sufficiently precise expression for so important a climatic element. As long as we have no direct measurements of the intensity of insolation, the record of the mean cloudiness in the different months is the only clue we have to the direct optical and thermal effects of solar radiation in any climate. The portion of sky covered with clouds, *i.e.*, the cloudiness, usually shows a considerable diurnal variation, and for this reason it is advisable to describe this element fully by giving also the monthly means of cloudiness for the different hours of observation, at least for a morning, an afternoon, and an evening hour. In addition to these data, the number of clear days (mean cloudiness 0 to 2), and of cloudy days (mean cloudiness over 8), should also always be given.

The direct registration of the duration of sunshine has been begun within the last few years by the use of the Campbell-Stokes and other sunshine recorders. This marks a decided advance in the measurement of one of the most important climatic elements. Since all stations do not have a sufficiently free horizon, and since the records when the sun is low are not so trustworthy for purposes of comparison as could be desired, it would be advisable to make a separate entry of the frequency of

sunshine between 8 or 9 A.M. and 3 or 4 P.M., besides the usual record of the total amount of sunshine. A record of the hourly frequency of sunshine is naturally to be especially recommended (see Table, p. 81, columns 8b and 8c).

Fog, which is cloud resting upon the earth's surface, is taken account of in climatology, not only because it checks insolation and nocturnal radiation, but also because it is a source of atmospheric humidity. Fog usually gives no measurable precipitation, but it may, nevertheless, to some extent replace rainfall in so far as the needs of vegetation are concerned. During dense fogs, the water which collects on trees and drops to the ground underneath them may have the same effect as a light rain in moistening the soil. Very considerable differences of temperature often occur within short distances during the prevalence of fogs. The *number of foggy days* is therefore one of the most important climatic factors. Besides this, a statement concerning the total number of hours of fog is desirable, so that it may be known whether the fog occurs only in the morning and evening, or whether, at certain seasons, it lasts all day.

Dew.—As in the case of fog, so also dew performs the function of rain during the dry season in some climates, and moistens the vegetation without being noticeable in the measured amounts of precipitation. The recorded rainfall, therefore, does not give a true measure of the amount of condensed water vapour which is available for the uses of vegetation.

Measurement of the quantity of dew.—No instrument which is generally available for the measurement of dew (drosometer) has yet been constructed. One of the chief difficulties in this matter is the fact that the quantity of dew deposited depends upon the nature of the body which is exposed to cooling by nocturnal radiation. If we wish to ascertain the amount of water which is condensed upon the leaves of plants, we should use these leaves themselves as drosometers. Moreover, the amount of dew which is deposited during the night is so slight in most climates that it is difficult to measure it before it evaporates. On the other hand, in some tropical countries, and even in valleys at considerable altitudes in our own climate, the dew is often so heavy on clear summer mornings that it may give the same amount of water as a light fall of rain. In dry continental interiors dew is unknown in some localities.

Instruments for the measurement of dew have lately been designed by Houdaille, of Montpellier (*M.Z.*, X., 1893, 433-434), and by Fr. von Kerner (*M.Z.*, IX., 1892, 106-108). The deposit of dew is very small in the dry climate

of Montpellier, where the average annual amount of dew as determined by three years' observations is 8·0 mm. The mean daily amount of dew deposited at Montpellier is about 0·08 mm. (*M.Z.*, XV., 1898, 72). Homén, as the result of his investigations in Finland, calculates the total depth of water deposited by a heavy dew upon a grass-covered surface as between 0·1 mm. and 0·2 mm. Dines, in England, puts it at barely 0·1 mm., and the total annual amount of dew at 26 mm. Wollny, in Munich, puts the total at 30 mm. (E. Wollny: "Untersuchungen über die Bildung und Menge des Thaues," *Forschungen auf dem Gebiete der Agrikulturphysik*, XV., 1892, III., 151. (*M.Z.*, IX., 1892, [93]-[94].). In mountains, and on coasts and islands within the tropics, the amount of dew is equal to that of a considerable rainfall. Pechuel Lösche estimates the amount of dew deposited during a favourable night on the Loango coast at 3 mm.; and there are many such nights during the dry season. Large puddles sometimes formed before midnight on the table which was used in these experiments, and which had been painted with green oil colours. ("Die Loango Expedition," 1873-76, III., 1882, 72.) Gräffe, writing of the Samoan Islands, says that the deposit of dew is so heavy during the dry season that it often results in a light rainfall in the woods, which wets the traveller to the skin in the early morning. (*Z.f.M.*, IX., 1874, 134-136.) A similar experience may also be met with in the Alps. The view which is now so generally held, that dew is of no benefit to vegetation, is probably as far from the truth as was the exaggerated view formerly held as to its importance. According to experiments made by Burt, the leaves of growing plants absorb water during the night, when they are damp, and are in a damp atmosphere: that is, when the conditions are those in which dew forms (E. A. Burt: "Do the Leaves of our ordinary Land-Plants absorb Water?" *Science*, XXII., 1893, 51-52.)

Dew often protects vegetation from freezing at night. The temperature of the leaves cannot fall below the dew-point of the air, because when the dew-point is reached, dew begins to form and further cooling is retarded by the resulting liberation of latent heat. The temperature falls lower, the drier the air; not because dry air is more diathermanous, but because its dew-point is lower.

Professor Willis L. Moore, Chief of the United States Weather Bureau, notes some interesting cases from Wisconsin which illustrate this point.[1] Horticulturists and growers of tobacco realise the great benefit which may be derived from the frost warnings of the Weather Bureau. As a result of investigation in connection with frost predictions, it appears that the conditions of soil-moisture must be taken into account, as well as the fall of temperature. When an area of high pressure approaches from the west, attended by clearer and colder weather, the most important element to be considered in determining the likelihood of occurrence of a frost that will injure the growing crops is whether or not rain has recently fallen. If the soil is dry, frost will

[1] "Relations of the Weather Bureau to the Science and Industry of the Country," *Science*, N.S., II., 1895, 576-582.

occur, but if a light rain of 10-12 mm. has recently fallen over a considerable area, there is no danger of frost. Moore finds that the minimum temperatures in cranberry marshes during abnormally dry seasons often fall 8° below the temperatures telegraphed from the cities and towns within a few miles of the marshes. The danger of damage by frost may be diminished by flooding the beds on the receipt of a frost warning, and during dry weather. It is evident that nocturnal radiation and a considerable reduction of temperature are checked by the formation of light fog sheets which are readily developed over moist soils.

CHAPTER III.

WINDS, PRESSURE, AND EVAPORATION.

Importance of wind as an element of climate.—Atmospheric currents are important climatic factors for many reasons. In the first place, the velocity of the air movement, apart from its direction, must be taken into account. The movement of the air increases evaporation and dries the soil, and thus increases the need of organisms for water and the evaporating capacity of a climate. This is but one aspect of the question; the other concerns the effect of the wind upon the temperature which is actually felt by the body, *i.e.*, the physiological temperature, which is not indicated by the thermometer. Winds usually have a cooling effect because they cause a more rapid conduction of heat from the body. A low temperature, which may easily be endured, and which is even agreeable when the air is calm, stimulating the body in many ways, becomes unendurable, or at least unpleasant, when the air is in motion. On the other hand, when the air is not very damp, very high temperatures are made much more bearable by a wind, because the latter increases evaporation. Such a wind may, however, be injurious to vegetation by quickly drying the more delicate portions of the plants. As a general rule, climates with a considerable air movement have a stimulating effect upon man, which is conducive to active work; while climates where calms prevail have an enervating effect, which is conducive to inactivity. The continual change of air when a wind is blowing is of marked hygienic importance in places where there is a large population crowded within narrow limits.

In the tropics, the stagnation of the air within enclosed valleys is often the cause of great unhealthiness. The prevalence of morning fogs in such localities is a proof of this, and it is best to avoid these

places, if possible, even in our climate. The southern hemisphere has, as a whole, better ventilation than the northern, in corresponding latitudes; and for this reason, and because of the greater dryness of the air, the former is, in general, more wholesome.[1]

Wind velocity: anemometers.—It would, therefore, be of distinct value if climates could be compared with one another on the basis of velocity of air movement, but, unfortunately, this is hardly possible as yet. Wind velocity is only *estimated* at most meteorological stations, a calm being denoted by 0, and a wind of hurricane velocity by 6, or 10 (at sea, 12 is used). The wind velocities which come between these limits are designated by the numbers which seem best to express the velocity in question. It is clear that the diurnal and the annual periods of wind velocity at any one station can be fairly well ascertained by this method. It is also clear that the estimates made at different stations are hardly comparable, for the reason that each observer will interpret the scale in accordance with the extremes of velocity which occur at his own station. Yet even when wind velocities are registered by means of anemometers, and are given in meters per second, these velocities are so much influenced by the accidental peculiarities of the local exposure of the anemometer that, in most cases, the recorded wind velocities are not those which really occur over the district as a whole. For this reason, estimates of wind velocity are sometimes to be preferred to actual measurements, because the former are based upon the observed effects over an extended area, while the latter hold for one fixed point only, which often does not, by any means, represent the average air movement. Further, there is the difficulty that even the best of the anemometers now in use, if they are of different sizes and forms, do not furnish data which are directly comparable. For these several reasons, comparable data concerning wind velocities in the different climates are not yet available. (In the table on page 81, column 9 gives the monthly means of wind velocity within the city; column 9*b* gives these means for the top of a tower 22 meters high in an open situation outside the city.) Estimates of wind velocity at sea, especially those made on board of sailing vessels, give results which, for obvious reasons, are much more comparable.[2]

[1] Cf. Radau: *Rôle des Vents dans les Climats chauds*, Paris, 1880; and Pauly: *Climats et Endémies*, 1874.

[2] Frank Waldo: "Distribution of Average Wind Velocities in the United States," *Am. Met. Journ.*, VI., 1889-90, 219-228; 257-266; 368-382; "The Geographical Distribution of the Maximum and Minimum Hourly Wind Velocities, and their Relations to the Average Daily Wind Velocities for January

Frequency of different wind directions.—The different wind directions must also be considered, from the point of view of their frequency, and of their special peculiarities, which often vary greatly. Thus the *frequency of winds from the different directions* becomes an important climatic factor. In most cases it suffices to record the frequency of winds from the eight principal points of the compass, for the results are then definite and can easily be compared. If it happens that a wind of one of the intermediate directions is particularly noticeable by reason of its constancy, or because of any special peculiarities, a record can be made of this, outside of the regular table reserved for wind observations.

The frequency of the different winds is most conveniently expressed in percentages of the total number of observations. Or else the number of observations of each wind is divided by the number of observations made each day, and thus is obtained the number of days that each wind has blown. These data should be given for every month, and not for the whole year only, because in most climates there is a more or less marked annual period of wind direction.[1]

In many localities, as on coasts and in valleys, the wind changes very regularly according to the time of day; and the direction of the wind in the morning is different from that in the afternoon. For this reason it is necessary, in such cases, to give the frequency of the winds for each hour of observation. In most cases, however, it suffices if this be given for longer periods than a month; namely, for a period of six months, or even of a year, for example, because the diurnal period does not vary essentially from month to month.

Wind roses.—The study of the influences of the varying frequencies of the different wind directions upon the climate of any place further necessitates some indication as to the meteorological peculiarities of the principal winds. This is accomplished by means of the so-called *wind roses.* The determination of the average temperature, humidity,

and July for the United States," *Ibid.*, XII., 1895-96, 75-89; "Synchronous or Simultaneous Geographical Distribution of Hourly Wind Velocities in the United States," *Ibid.*, 145-151; "Wind as a Motive Power in the United States," *Review of Reviews*, New York, XII., 1895, 299-302.

S. P. Fergusson: "Anemometer Comparisons," *Annals Astron. Obs. Harv. Coll.*, XL., Part IV., Appendix G, 4to, Cambridge, Mass., pp. 35, pls. 3. R. H. Curtis: "An Attempt to Determine the Velocity Equivalents of Wind Forces estimated by Beaufort's Scale," *Quart. Journ. Roy. Met. Soc.*, XXIII., 1897, 24-41 [*M.Z.*, XIV., 1897, (51)-(53).]

[1] The wind table for Vienna has been omitted for lack of space.

cloudiness, and rain probability, which are associated with each wind direction enables us to draw wind roses that illustrate these several relations. Since the general meteorological characteristics of the principal wind directions are very uniform over considerable areas, the climatic features of a large section of country can be sufficiently described by means of one wind rose, or at most by means of a few such figures. Local exceptions certainly occur, but the meteorological effects of these special local winds, such as the foehn, the sirocco, or the bora, can easily be considered by themselves.

Control of weather by winds.—In the belts which lie between the limits of the torrid zone and the inner portions of the polar regions the control of climate is really determined by the winds. The weather conditions are often completely reversed when the wind changes from one direction to another, the same conditions continuing as long as the wind blows from the same quarter. When northeast and east winds blow for a considerable length of time in western Europe they cause an invasion by the dry, clear, continental climate, which is cold in winter and warm in summer. When, on the other hand, southwest and west winds prevail, the influence of the ocean extends far inland, the weather is damp and cloudy, being warm in winter and cool in summer. Thus the winds are seen to obliterate climatic barriers, and to keep neighbouring climates in a constant state of interchanging conditions. There are but few districts which may be said to have their own climates. Among these are included those that are shut off from their surroundings on all sides by high mountains, like eastern Turkestan and eastern Siberia in winter. In almost all cases, changes of weather are produced by a continuous displacement of climatic boundaries by the winds. Especially is this true of those districts which lie between two well-marked controlling climatic types, and of middle latitudes, as has been pointed out. In the torrid zone, and in the interior of the polar regions, the winds lose this marked control over the weather, as is shown by the meteorological wind roses, which make it clear that the differences between the average values of the meteorological elements associated with different wind directions are very small and unimportant.

Pressure as a climatic factor.—Atmospheric pressure and its variations are of secondary importance as climatic factors; a condition of things which is in strong contrast with the important part which these elements play in meteorology. When the climates of individual stations are to be described, observations of pressure can be completely discarded. In so far as pressure has any influence upon organic life, this factor is

known with sufficient accuracy when the altitude of the station above sea-level is given. Indeed, the pressure can be obtained with greater accuracy by calculation than by derivation from barometric observations which are carelessly made. The pressure on the fourth floor of a city house is about 2 mm. less than on the ground floor, and there are much greater differences of pressure in many cities which are built on uneven ground, without there being any noticeable differences of climate in consequence. Hence it may be inferred that, in a simple description of a climate, the pressure need not be known within less than a centimeter.

The variations in pressure at the same place likewise hardly need to be considered in climatology. Their influence is, to say the least, easily over-estimated. In most places the variations in pressure are much too slight to have any influence upon organic life. Changes in pressure of 20 mm. in 24 hours seldom occur, and then only in certain regions. The effects of such a change upon the human body can be imagined when we reflect that we should experience the same change if we were very gradually carried from the bottom to the top of a hill only 200 meters high during a day. It is hardly to be supposed that this could have any noticeable physiological effect.[1]

Relation of pressure to evaporation.—From a climatological point of view the pressure which prevails at any place is considered not as pressure only, but also because it is an indication of the rarity of the air, and especially because of its influence upon evaporation. As is well known, the decrease of atmospheric pressure increases evaporation, temperature, air movement, and relative humidity remaining the same.[2] For this reason evaporation is, in general, considerably greater in lofty valleys and on mountains than over lowlands.

An approximate knowledge of the pressure is all that is necessary in the various relations thus far considered. When, however, we study the interrelations of different climatic districts as controlled by the winds, an accurate knowledge of the pressure becomes necessary.

[1] Changes in pressure have no injurious effects upon health. In experiments with pneumatic chambers, pressure changes amounting to 300 mm. a day have been produced without causing any notable injurious effects upon the sick persons concerned in these experiments (Thomas : *Beiträge zur Allgemeinen Klimatologie*, Erlangen, 1872).

[2] According to Stefan, the rate of evaporation is proportional to the logarithm of a fraction whose numerator is the air pressure, and whose denominator is the air pressure diminished by the maximum vapour-tension.

The altitude of the station at which the observations are made must also be accurately known, because near sea-level, for example, a difference of elevation of about 10 meters involves a difference of pressure of 1 mm.[1] In theoretical climatology, which studies the causes of the distribution of the different climates, it is of the greatest importance that the monthly means of atmospheric pressure should be known at a large number of stations, distributed as uniformly as possible over the earth's surface, because the prevailing winds depend upon the distribution of pressure, and the winds closely affect the temperature and the humidity of the air. It is, however, necessary to emphasise the fact that detailed statements concerning the pressure conditions of any single station give no information as to its climate, and that pressure data have no value until they can be compared with similar data for other stations at the same altitude. Even then they give no direct information concerning the climate itself; they furnish only a basis for the explanation of the distribution of the other climatic factors.

Evaporation.—The total effect of air temperature, pressure, relative humidity, and average wind velocity upon a free water surface in the shade or in the sun is expressed by the amount of water evaporated.[2] The measurement of evaporation would be of great importance in climatology, because it would also give an approximate indication of the need of organisms for water in each climate. It is, unfortunately, a difficult matter to make measurements of evaporation which shall be strictly comparable. In order to carry out such measurements it would be advisable to use evaporimeters which are precisely alike, and to expose them all in exactly the same way.

[1] In this connection see "The Effects of Diminished Pressure on Cooking," *Mo. Weather Rev.*, XXVIII., 1900, 160-161.

[2] A. Weilenmann: "Berechnung der Verdunstung aus den meteorologischen Faktoren," *Z.f.M.*, XII., 1877, 268-271; 368. E. Wollny: "Untersuchungen über die Verdunstung," *Forsch. a. d. Geb. d. Agrikulturphysik*, XVIII., 1896, 416-486 (*M.Z.*, XIII., 1896, 362-364). For evaporation data concerning the United States see D. Fitzgerald: "Evaporation," *Trans. Am. Soc. Civ. Engin.*, XV., 1886, and T. Russell: "Depth of Evaporation in the United States," *Mo. Weather Rev.*, XVI., 1888, 235-239.

Fitzgerald found the annual evaporation at the reservoir of the Boston, Mass., waterworks 993·4 mm. Russell computed the annual evaporation in shelters of the Signal Service Stations from means of observations of dew-point and wet-bulb taken thrice daily. The maximum evaporation thus computed was 2570·5 mm., at Fort Grant, Ariz., and the minimum, 459·7 mm., at Tatoosh Island, Wash.

Relative values are alone of importance; absolute values depend upon so many accidental circumstances that we cannot expect to obtain them. Even the amount of evaporation from a perfectly free water surface under sunshine is uncertain, because this depends also upon the depth, extent, and temperature of the body of water, and upon many other local conditions. The Wild evaporimeter is the best.[1]

[1] Concerning the diurnal march of evaporation, see Houdaille, *M.Z.*, X., 1893, 431-432. Woeikof also discusses evaporation as a climatic factor (*Die Klimate der Erde*, I., 19, *et sqq.*).

CHAPTER IV.

THE COMPOSITION OF THE ATMOSPHERE.

The composition of dry atmospheric air in widely separated portions of the earth's surface is so extraordinarily uniform, that the differences which have been discovered are usually smaller than the recognised limit of error in the analysis. Dry atmospheric air is well known to be a mixture of 21 parts of oxygen and 79 parts of nitrogen, by volume, in addition to which there is a very small percentage of carbon dioxide, about 0·03 per cent., or 3 parts by volume in 10,000 parts. In our ordinary damp air, water vapour is also present in very variable amounts. The vapour tension, which has already been referred to as a climatic factor, gives a simple measure of the quantity of water in the air. If the vapour tension is divided by the air pressure, the quotient is the percentage of vapour in the air by volume. In damp countries near the equator, as at Batavia, for example, where the vapour pressure is 21 mm., the amount of water in the air rises to nearly 3 per cent. of the whole, by volume,[1] while in central Europe, even in summer, with a vapour pressure of about 10 mm., it amounts to only 1·3 per cent., by volume. The differences in the composition of the atmosphere in different climates, so far as these differences can be directly proved, therefore really concern water vapour only. Variations in the amount of carbon dioxide need hardly be considered at all.

[1] More exactly, 2100 : 760 = 2·8 per cent. The composition of the atmosphere under these conditions is as follows :—76·8 per cent. of nitrogen ; 20·4 per cent. of oxygen ; and 2·8 per cent. of water vapour, by volume. In addition, there are a few hundredths of a per cent. of carbon dioxide. The greatest vapour pressure known to the author is 30·7 mm., observed once during the rainy season at Allahabad. This is equal to four per cent. by volume. The very small amounts of argon, metargon, krypton, and neon which have

Water vapour and respiration.—Dr. Ucke has calculated the effect of the amount of water vapour in the atmosphere in different climates upon the absorption of oxygen during the process of respiration in man. Water vapour acts as a diluter of the air. Ucke finds that in normal, average respiration, a man in the damp tropical climate of Madras takes in 80·7 kg. of oxygen every month. In London and Brussels, on the other hand, the amount is 87·3 kg.; and in the still drier climate of St. Petersburg and Barnaul, it is 90·4 to 90·7 kg. In the extremely cold winter air of Barnaul this amount may even increase to 99·2 kg. in January. It must, however, be added that even a slight increase in the altitude of a station above sea level, with the accompanying decrease in pressure, has the same effect in this respect as the greatest humidity of the air in the tropics. Thus Ucke finds that the amount of oxygen breathed in on the Peissenberg in Bavaria, at an altitude of 1000 m. above sea level, is 79·2 kg., which is still less than in Madras. No important climatic influence can therefore be claimed for this effect of water vapour.

Carbon dioxide.—The amount of carbon dioxide in the atmosphere in different climates has frequently been determined. In central Europe, the recent analyses of the air, which have been regularly and frequently made, show that the carbon dioxide amounts to 0·03 per cent., by volume, while it was formerly assumed, on the basis of older analyses, that the amount was 0·04 per cent. Saussure, in the case of the Alps, and especially the Brothers Schlagintweit in the case of the Alps and of the Himalayas, believed that they could prove an increase in the proportion of carbon dioxide with increasing altitude. The accurate analyses of Müntz and Aubin, however, showed but 0·028 per cent. at a height of 2877 m. on the Pic du Midi, which is the same amount as at 600 m. at the base of the mountain. Further recent analyses of air from Grands Mulets (3050 m.) show a very slight decrease in the proportion of CO_2 with increasing altitude.[1] The analyses of samples of air collected at great heights by the aid of balloons show no change in the amount of CO_2 in the air with increasing altitude.[2]

An analysis of a sample of air obtained at an altitude of 15,500 meters, during the ascent of the *ballon-sonde* "L'Aerophile" on Feb. 18, 1897, gave, according to Müntz, 0·033 per cent. of CO_2.[3]

recently been discovered in atmospheric air are not taken into consideration here. Argon is a gas which is very much like nitrogen, and somewhat heavier. It has only recently become possible to separate the argon from the nitrogen See Rayleigh and Ramsay; "Argon, a new Constituent of the Atmosphere," *Smiths. Contr. to Knowl.*, 1033, fol., Wash., D.C., 1896, pp. 43. W. Ramsay: *The Gases of the Atmosphere: The History of their Discovery*, 8vo, London, 1896, pp. viii. + 240.

[1] *Comptes Rendus*, CXXIX., 1899, 315 (*M.Z.*, XVII., 1900, 87-88).

[2] *M.Z.*, XIII., 1896, 144-145.

[3] *Comptes Rendus*, CXXIV., 1897, 486-488 (*M.Z.*, XIV., 1897, 155-156).

Further, Ebermayer has shown that the amount of carbon dioxide in the air of forests does not differ materially from that in the air of the open country.[1]

The proportion of carbon dioxide in the atmosphere is somewhat greater by night than by day. The southern hemisphere seems to have a somewhat smaller proportion of carbon dioxide than the northern, which probably results from the large extent of water surface, and from the lower temperature, in the former hemisphere. This same result has also been reached by Schlösing in a theoretical calculation based upon the mutual relations of the carbon dioxide in the water and in the atmosphere.

The proportion of carbon dioxide in cities is somewhat greater than in the country. Müntz and Aubin found 0·031 per cent. in Paris, although this amount was subject to variations. In the neighbourhood of Paris they found 0·028 per cent. The proportion of carbon dioxide increases during fogs. In London, Russell found 0·038 per cent. in clear weather, 0·045 per cent. in cloudy weather, and 0·051 per cent. during fog. The maximum noted during fog was 0·14 per cent.

Oxygen.—The proportion of oxygen in the air of Munich was found by Jolly to vary between 21 per cent. and 20·5 per cent., by volume. The smallest proportion of oxygen was noted with south, southwest and west winds; and the highest with north and northeast winds. Similarly, Macagno, at Palermo, found the proportion of oxygen smaller during a sirocco (south wind), namely, about 20 per cent., as against a mean of 20·8 per cent., based upon numerous analyses. The proportion of oxygen in the air as a whole is, however, very constant. Numerous analyses of air made from Tromsö to the equator have shown no notable differences in the relative amounts of the oxygen. The percentages are as follows: Tromsö, 20·92; Dresden, 20·90; Bonn, 20·92; Cleveland, 20·93; Para, 20·89. The absolute extremes in 203 analyses were 21 per cent. at Tromsö, and 20·86 per cent. at Para.[2] The air collected at great altitudes during balloon ascents has also shown hardly any decrease in the proportion of oxygen.

Dust, smoke and other impurities.—Of the various local components, or impurities, of the air which are of a coarser character, dust and even

[1] Ebermayer: *Die Beschaffenheit der Waldluft.* Stuttgart, 1885. For further references to investigations concerning the amount of CO_2 in the atmosphere see E. A. Letts and R. F. Blake: "The Carbonic Anhydride of the Atmosphere," *Sci. Proc. Roy. Dublin Soc.*, IX., N.S., March, 1900, 107-270. This paper contains a comprehensive bibliography.

[2] W. Hempel: "Ueber den Sauerstoffgehalt der atmosphärischen Luft," *Bericht Deutsch. Chem. Gesell.*, XX, 1887, 1864-1873.

smoke need consideration from a climatic point of view. Dust is found very generally in the air, especially in the dry interiors of Asia, Africa, and Australia. Smoke occurs only occasionally. Forest or prairie fires produce smoke, and smoke also plays a part in the formation of "city fogs." The pollution of the atmosphere by the smoke over large cities, and the formation of fog, which goes hand in hand with it, have gradually come to be a more and more serious injury to people who live in cities. The unwholesomeness of the smoky atmosphere of cities is due not only to the diminished intensity of daylight, and to the shutting out of the health-giving sunshine, which acts as a disinfectant by destroying most of the harmful germs; it is due also to the actual pollution of the air by means of the sulphurous gases and organic substances which accumulate within the foggy strata. The air in the town hall of Manchester, England, was found to contain 9·3 mg. of sulphuric acid (SO_3) in every 100 cubic feet in winter, and 2·8 mg. in summer. The maxima, in December and March, were 21 and 27 mg. W. J. Russell has investigated the impurities in the air of London, and also the origin and effects of city fogs.[1] He gives the analyses made of fog deposits which occurred in February, 1891, in and near the city of London. The deposits were obtained from the previously washed glass roofs of the plant-houses at Kew and from some orchid-houses at Chelsea. At Kew, 20 square yards of roof yielded 30 grams of deposit, and at Chelsea the same area gave 40 grams, which represents 22 lbs. to the acre, or 6 tons to the square mile. The composition of the deposits was as follows:—

	Chelsea. Per cent.	Kew. Per cent.
Carbon,	39·0	42·5
Hydrocarbons,	12·3	4·8
Organic bases (pyridines, etc.),	2·0	
Sulphuric acid (SO_3)	4·3	4·0
Hydrochloric acid (HCl),	1·4	0·8
Ammonia,	1·4	1·1
Metallic iron and magnetic oxide of iron,	2·6	41·5
Mineral matter (chiefly silica and ferric oxide),	31·2	
Water, not determined (say difference),	5·8	5·3
	100·0	100·0

[1] "Impurities in London Air," *Monthly Weather Record*, August, 1885. Also, F. A. R. Russell: "Smoke in Relation to Fogs in London," *Nature*, XXXIX., 1888-1889, 34-36 (*M.Z.*, VI., 1889, 33-35); W. J. Russell: "Town Fogs and their Effects," *Nature*, XLV., 1891-92, 10-16 (*M.Z.*, IX., 1892, 12-21); F. J. Brodie: "On the Prevalence of Fog in London during the Twenty years 1871 to 1890, *Quart. Jour. Roy. Met. Soc.*, XVIII., 1892, 40-43; J. B. Cohen, "The Air of Towns," *Smithsonian Miscellaneous Coll.*, 1073, 8vo, Washington, D.C., 1896, pp. 41, pls. 21.

Professor Thistelton Dyer said of the deposit collected at Kew that "It was like a brown paint: it would not wash off with water, and could only be scraped off with a knife."

Aitken's interesting studies concerning the number of dust particles in the air at different places hardly come within the field of climatology.[1]

London fogs.—The development of black city fogs, especially of London fogs, is thus described by Russell: "An ordinary thick white fog covers the city, say at 6 A.M.: about a million fires are lighted soon after this hour, and the atmosphere becomes charged with enormous volumes of smoke — that is, the gases of combustion, bearing carbonaceous particles. Now, these particles, as soon as they are cooled to the temperature of the air, or below it, begin to attach to themselves the water spherules already present and visible, and vapour may also be condensed on the particles. A thick layer of these united particles prevents the light from penetrating it, and a very small quantity of finely divided carbon may stop the bright sunshine altogether, like the film of soot on a smoked glass. . . . Smoke prevents the warmth of the oblique sunshine from reaching and evaporating the white fog near the ground, and the white fog continues to radiate towards space and towards the ground, if colder than itself, without receiving compensation from the solar rays. . . . Carbon is a good radiator, and tends from this cause to keep itself cold by radiation into space, and thus to accumulate vapour from the air, like the dewy surface of the earth. . . . On a fine, cold, still morning, with a bright sun and temperature near the dew-point, persons arriving from the country are pretty sure to find a black fog in town between 10 and 12 A.M." The city of London has, on the average, but 15 hours of sunshine during each of the months of December and January. Greenwich, which is close to the city but outside of it, has 45 hours of sunshine; and Kew has 71 during each of these months. The yearly amounts are 1026, 1227, and 1399 hours respectively. This deficiency in the amount of light must have a very injurious effect upon the vitality of the inhabitants of cities. During the great fog in the winter of 1880, the excess of deaths during three weeks in London (January 24 to February 14) was 2994; and the excess of cases of sickness was probably ten times greater.

Brodie gives the following table showing the increase in the number of fogs in London during five-year periods from 1871 to 1890:—

NUMBER OF DAYS OF FOG IN LONDON IN FIVE-YEAR PERIODS.

1871-1875	1876-1880	1881-1885	1886-1890
50·8	58·4	62·2	74·2

[1] See papers by John Aitken in *Trans. Roy. Soc.*, Edinburgh, XXX., 1880-83, 337; XXXV., 1887-90, 1-19; 337-368; XXXVI., 1889-91, 313-320; XXXVII., 1891-1895, 413-425; XXXVIII., 1891-1895, 17-50; 621-694. *Proc. Roy. Soc.*, Edinburgh, XVI., 1888-89, 135-172; XVII., 1889-90, 193-254; XVIII., 1890-91, 39-52; 259-262; XIX., 1891-92, 260-263; XX., 1892-95, 76-93. Abstracts or

Micro-organisms.—There are also other impurities in the atmosphere which are of a much more subtile character; these are not perceptible by the senses, and their presence cannot be detected even by chemical analyses. Nevertheless, these very impurities are of the greatest climatic importance because of their injurious effects upon the human body.[1] These disease-bringing impurities, which are mostly confined to the lower strata of the atmosphere, were formerly known as miasmas. These miasmas were supposed to be noxious gases, but they have recently been shown to be organic germs which, when taken into the body, may further develop and increase in number, causing sickness, and even death. Malaria is the greatest scourge of warm and hot climates, and also occurs sporadically in cooler climates, although progress in the cultivation and draining of the soil, as well as improved dwellings and conditions of life, have very markedly narrowed its limits within the cooler climates. Malaria is not found in dry climates, nor in deserts and on steppes. High latitudes, and high altitudes upon mountains, even within the tropics, are free from malaria.

In 1879, Tommassi-Crudeli and Klebs discovered, in the air and in the soil of the malarial districts of Italy, a bacillus which they called *Schizomycetes bacillaris*, and which, after being isolated and cultivated, produced marked symptoms of malaria in animals into whose blood it was introduced. Laveran was the first to find the malaria bacillus in the blood of persons ill with malarial fever. Recent investigations by Manson, Koch, Ross and others, have led to the belief that malaria is carried by a certain genus of mosquito (*anopheles*), and that infection results from the bite of mosquitoes which have previously been infected.

Pure air, especially such as is notably free from organic impurities, is found over the ocean, and on sea coasts when the wind is blowing off the water. Ocean air contains some salt[2] and traces of iodine, which may be detected even by their odour. Desert air, also, is pure and wholesome, as is the air of elevated mountainous districts. The latter is at the same time rarer, by reason of the altitude. Rain cleans the air; it removes the impurities for a time, at least, and brings down air which is fresh and pure from the upper strata of the atmosphere. The

reprints of some of these papers will be found in *Nature*, XXXVII., 1887-88, 428-430; XLI., 1889-90, 394-396; XLIV., 1891, 279; XLV., 1892, 299-301; 582-584; XLIX., 1893-94, 544-546.

See also J. R. Plumandon: *Les Poussières atmosphériques. Leur Circulation dans l'Atmosphère et leur Influence sur la Santé*, 8vo, Paris, 1897, pp. 130.

[1] D. H. Bergey: "Methods for the Determination of Organic Matter in the Air," *Smiths. Contr. to Knowledge*, 1037, 8vo, Washington, D.C., 1896, pp. 28.

[2] Verhaege found 0·2 grams of salt in 2000 liters of sea air.

latter effect, which is produced by heavy showers of rain, has but recently been appreciated. The air which is driven down and carried along by the rain, makes itself felt as a gusty wind blowing out in front of, and beneath, the rain cloud.

Ozone and other constituents.—The climatic importance of ozone is a somewhat debatable question, but it cannot be doubted that the presence of ozone in air shows that this air has active oxidising properties, whether this fact is to be ascribed to the more active form of oxygen which is called ozone, or to the presence of peroxide of hydrogen. When ozone is present in considerable quantity in the atmosphere, it is a sign that the air is free from organic impurities and products of decay. Inhabited rooms and vitiated air show no ozone reaction. The ordinary measurements of the amount of ozone in the air, by means of the so-called ozonometer, do not permit of any rigid comparisons or general conclusions. (The result of the ozone measurements made at Vienna according to Schönbein's scale of 1-10 are given in columns 11 and 12 of the table on page 81.) The average amount of ozone in 100 cubic meters of air is 1·4 mg. according to measurements made at Montsouris (Paris). The amount of ozone may be doubled after thunderstorms, as was shown by Schönbein, who found 2·6 mg. under those conditions. Analyses of air from Chamonix and from the Grands Mulets on August 23 and 24, and Sept. 4, 1896, showed, according to de Thierry, an increase of the ozone content with increasing altitude. The amount of ozone in the air at the Grands Mulets was found to be about four times as large as that in the air of Paris.[1] At the Montsouris Observatory, regular observations are also made of the amount of carbon dioxide and ammonia in the air (2·0 mg.), as well as of the number of bacteria in a cubic meter, and of the amount of ammonia and nitric acid in rain. The average number of bacteria in a cubic meter of air at Montsouris has been found to be 345, while in the interior of Paris it is 4790. In the annual reports of the Montsouris Observatory the results of all these different investigations are given in detail, together with the average values derived from the whole period of observation. Miquel has also determined the diurnal range of the numbers of bacteria in the air of Paris.[2]

[1] M. de Thierry: "Dosage de l'Ozone atmosphérique au Mont Blanc." *Comptes Rendus*, CXXIV., 1897, 460-463 (*M.Z.*, XIV., 1897, 155).

[2] In *M.Z.*, IX., 1892, 101-103, there is an abstract from the volume for 1891 of the *Annuaire* of the Montsouris Observatory, with data concerning the annual periodicity of the amounts of ozone, ammonia, carbon dioxide and bacteria in the air of Paris.

TABLE II.

CLIMATIC DATA FOR VIENNA—CONTINUED.

B. HUMIDITY, PRECIPITATION, CLOUDINESS, WIND VELOCITY, EVAPORATION, OZONE.

	Humidity (20 years).					Rain and Snow.					Mean Wind Velocity (meters per second).			Ozone 0 – 10 (20 years).	
	Mean Vapour Tension. (mm.)	Relative Humidity (in percentages). 6 a.m.	2 p.m.	10 p.m.	True Mean.	Amount (34 years). (mm.)	No. of Rainy Days (20 years).	Cloudiness 0 – 10 (20 years).	Mean Duration of Sunshine in Hours. (15 years).	Ditto in Percentages of greatest possible Duration. (15 years).	City (6 years).	Observatory (20 years).	Evaporation (5 years) (mm.)	Day.	Night.
December,	3·7	86	77	86	83	40	12·4	7·3	49·1	19	2·4	5·2	18	3·1	5·5
January,	3·6	87	77	86	84	35	12·8	7·2	65·5	23	1·7	5·1	13	3·2	5·8
February,	3·8	84	70	83	80	36	11·2	6·7	86·5	30	2·6	5·4	27	4·2	6·0
March,	4·4	81	58	76	71	43	13·1	6·2	129·8	36	2·2	6·2	39	4·2	6·2
April,	5·7	76	48	68	63	42	12·3	5·2	178·9	43	2·4	5·2	71	4·6	5·7
May,	8·2	76	49	71	64	64	13·0	5·1	240·9	51	2·0	5·4	87	5·2	5·4
June,	10·0	75	50	71	64	66	12·7	4·9	228·2	48	2·4	5·3	93	5·2	5·6
July,	10·9	75	48	70	63	65	13·3	4·5	274·0	56	2·2	5·5	113	5·3	5·3
August,	11·0	79	50	73	66	72	11·8	4·5	246·1	56	2·1	4·9	94	5·1	5·4
September,	9·3	82	53	75	69	45	8·3	4·5	175·9	48	2·0	4·7	77	3·9	4·5
October,	7·2	85	61	81	76	44	10·6	5·4	100·5	30	2·0	4·6	47	3·0	4·3
November,	4·8	84	72	83	80	43	12·6	7·4	63·2	23	3·0	4·9	32	3·0	5·1
Year,	6·9	80	59	77	72	595	144·1	5·7	1838·6	41·4	2·2	5·2	711	4·2	5·4
Column,	1	2	3	4	5	6	7	8	8 b	8 c	9	9 b	10	11	12

There can hardly be any doubt that there are still other modifications in the composition of the atmosphere which have some influence upon the human organism, but which have not yet received any attention. When we recall what an enormous quantity of air a man breathes every day (about 10,000 liters) we should not be surprised to find that even very minute traces of certain substances, when mixed with the air, may be of importance to health; especially in view of the fact that these effects may be cumulative from one day to the next, if the body remains in such air for any length of time. Hardly any beginning has yet been made towards studying the atmosphere with this point in mind.[1]

Atmospheric electricity.—There is a general disposition to ascribe some influence upon organic life to the electrical condition of the air. This is, unhappily, still quite an untrodden field, and no definite results have been obtained. It would, nevertheless, be premature to deny that any such influence exists. Yet even if such an influence were clearly established, climatology would hardly be in a position to consider atmospheric electricity as a climatic factor. The observations hitherto made enable us to draw some conclusions concerning the diurnal and the annual march of atmospheric electricity at some few places, but a comparison of the electrical tension of the atmosphere in different climates is not yet possible. So far, there is no indication that atmospheric electricity plays a notable part in climatology.

Subjective temperatures.—In the enumeration of the climatic factors in the first edition of this book, Professor Abbe noted the absence of some points which will be briefly referred to here. Professor Abbe is of the opinion[2] that, in describing a climate, it is not sufficient to give only the number of cold and hot, of dry and wet days, etc., but that it is also necessary to classify and to record the number of days on which the sensations experienced by the observer are defined by such expressions as "harshness," "rawness," "penetrating,"

[1] F. A. R. Russell: "The Atmosphere in Relation to Human Life and Health,' *Smiths. Misc. Coll.*, 1072, 8vo, Washington, D.C., 1896, pp. 148, also 8vo, London, 1897.

The following books concern the composition and the impurities of the atmosphere, with special reference to hygiene:—Renk: *Die Luft*, in *Handbuch der Hygiene*, by Pettenkofer and Ziemssen, Leipzig, Vogel, 1886. Miquel: *Les Organismes vivants de l'Atmosphère*, Paris, Gauthier Villars. Miquel: *Die Mikroörganismen der Luft*, Munich, Prieger, 1889. G. Tissandier: *Les Poussières de l'Air*, Paris, Gauthier Villars. Additional references have already been given in this chapter.

[2] *Scientific Record for* 1883. Report on Meteorology, Annual Report of the Smithsonian Institution for 1883, 491.

"mildness," "softness," "exhilarating," "stimulating," "invigorating," "oppression," "weakening," "cheerfulness." This suggestion was likewise made by J. W. Osborne, who proposed a definite scale to be used in such observations. This scale is as follows:—20, Intolerably hot; 19, excessively hot; 18, very hot; 17, tolerably hot; 16, very warm; 15, decidedly warm; 14, agreeably warm; 13, mild and soft; 12, mild and fresh; 11, quite fresh; 10, very fresh; 9, decidedly cool; 8, very cool; 7, moderately cool; 6, cold and fine; 5, cold and sharp; 4, very cold; 3, bitterly cold; 2, painfully cold; 1, unbearably cold. Observations made according to this plan were to be recorded four times a day, and were to be sent in every week on a postal card. It was believed that the results would be of service to biologists and physiologists in studies of the effects of different temperatures.[1] Unfortunately, it would be somewhat difficult to have the observations made according to any such scale at all comparable. The uncertainty as to the meanings of the different terms, and the influence of the observer's personal feelings in deciding among them, would be serious obstacles in the way of carrying out this proposal, which is otherwise certainly very worthy of attention.

Abbe also lays great weight upon the determination of the number of storm centres that pass over a given locality, or the storm frequency, and upon the direction of movements of these storms. We fully agree with Abbe in believing that charts of the tracks of barometric minima, and of their frequency of occurrence upon these tracks, are a valuable aid in descriptive climatology.[2] Such charts furnish direct evidence concerning the changeableness and the peculiarities of the weather at any definite place. Charts showing the distribution of storm frequency are therefore also of importance in determining the boundaries between climates. The frequency of weather changes from hot to cold and from cold to hot, as well as those from wet to dry, and *vice versa*, is an important climatic factor. Abbe also recommends notes on the appearance of the sky; whether it is clear and deep blue, or pale and dull. This is especially important in the tropics, where so much light and heat are distributed by diffusion that probably one-half of the total radiation of light and heat becomes effective in this way.

[1] J. W. Osborne: "Determinations of Subjective Temperature," *Proc. Am. Assoc. Adv. Sci.*, XXV., 1876, 66-74.

[2] A chart of this kind for the United States, prepared by Abbe, was published in the *Statistical Atlas*, Ninth Census, Washington, D.C., 1874. Köppen and van Bebber have charted the cyclonic tracks in Western Europe. See W. Köppen: "Erläuterungen zur Karte der Häufigkeit und der mittleren Zugstrassen barometrischer Minima zwischen Felsengebirge und Ural," *Z.f.M.*, XVII., 1882, 257-267. W. J. van Bebber: "Die Zugstrassen der barometrischen Minima," *M.Z.*, VIII., 1891, 361-366. Also Bartholomew's *Atlas of Meteorology*, pl. 28.

CHAPTER V.

PHENOLOGICAL OBSERVATIONS.

Phenological observations as climatic factors.—Before concluding the consideration of the climatic factors, a few words may be added concerning the use of phenological observations in climatological discussions. It has thus far been impossible to establish the dependence of the development of plant life upon temperature so securely as to allow of inferring the temperature from the observed stage of plant development, and *vice versa*. There is also but little prospect that it will be possible to use the phenomena of plant life as a trustworthy scale of temperature, because the physiological powers of adaptation to climatic peculiarities which the plants possess come into play, and these powers differ very much in different species of plants.[1] It is because of this power of adaptation that the same plant, when in different climatic surroundings, needs different amounts of heat and of humidity for the attainment of given stages of its growth. As Drude points out, "The birches at North Cape put forth leaves at lower temperatures than do similar trees on the Dresden Heath, and all their other stages of growth are accomplished in less time, and at lower temperatures. Beeches which have been introduced into Madeira shed their leaves at temperatures which keep our beeches still fully covered with foliage." Thus plants have a somewhat different scale of temperature for each locality. Notwithstanding this fact, it is best to take advantage of the assistance which may be gained from phenological observations in the discussion of local climatic conditions over a restricted area. Phenological obser-

Milton Whitney: "Meteorology as Distinguished from Climatology," *Science*, N.S., VII., 1898, 113-115; C. Abbe, *Monthly Weather Rev.*, XXVI., 1898, 168; Milton Whitney: "Climatology versus Meteorology," *ibid.*, 301-302.

vations, which are so readily made without the use of instruments, and with the expenditure of but little time, may at least serve as indications of existing climatic differences where regular meteorological observations are not to be expected. This is especially the case in mountainous districts, where considerable differences occur within a restricted area, and where the varying exposures of the mountain sides with reference to sunshine and warm winds play so important a part. The author entirely agrees with the following words of Professor Drude:—[1]

"In spite of the capacity for acclimatisation, there still remains a considerable difference in the time at which a given stage of growth is entered upon by a plant in different climates. With increasing latitude and altitude, there is always a retardation of the stages of development in the same plant. This retardation, when stated in *days*, may express the climatic differences of two places which are being compared more clearly than do the mean temperatures of these places, especially as agriculture in its different phases is associated with certain critical stages in the development of the wild plants, and not with certain temperatures. Observations of the time at which the same stages in the development of given plants occur at different places within small areas may give a clear and intelligible indication concerning the availability of the soil for cultivation."

It cannot be denied that the results of phenological observations do characterise differences in climate in a way which is easily understood.

By means of data of this sort we learn, for example, that the time of blossoming in early spring comes 20 to 25 days earlier in Triest, Görz, and Villa Carlotta than in Vienna; in Paris it comes 9 days earlier, while it is 16 days later in Lemberg, and 21 days later in Zlozow. We learn further that on the plateau of the Erz Mountains the retardation is just as great as at Moscow (34 days), and that the retardation in the Alps, in latitude 46° 30 N., and at 1700 meters above sea level, is 45 days, which is as great as at St. Petersburg. In order to emphasise existing climatic differences, especially within small areas, it is well to use the results of phenological observations, and not, by any means, to discard them. The results here referred to are the average dates upon which certain varieties of plants enter upon certain stages in their development.

Results of phenological observations.—Students of phenology have been especially occupied with the question as to the amount, and

[1] "Anleitung zu phytophänologischen Beobachtungen in der Flora von Sachsen," *Abhandl. Naturwiss. Gesellsch. Isis*, Dresden, 1881, 3-24. These instructions are to be highly recommended to those who wish to undertake such observations.

also the duration, of the heat which different plants need in order to attain certain stages in their development. At first a simple calculation was made which gave the total temperature that had accumulated from the beginning of the year up to the time of development of the special stage of growth in question. Temperatures below freezing were omitted in this consideration, because these correspond to a stagnation in the life of the plant, and therefore should not be taken into account in obtaining the final result. This method was chiefly employed by Fritsch, who used the greatest care in determining the phenological constants of temperature. It was, however, discovered that the accumulated temperature for the same stage of development of the same variety of plant varies a good deal from one year to the next. Hoffmann thought it better to take account, not of the shade temperatures, but of the maximum temperatures in the sun, and to deal with these latter in the same manner. He believed that these accumulated maximum temperatures would show more agreement. The "temperatures in the sun," it will be remembered, hold only for a given thermometer, with a given exposure, and a comparison of these temperatures is probably out of the question.

Besides the problem whether shade temperatures or temperatures in the sun should be employed, came the further question, whether the same starting-point in adding up the temperatures should be adopted for all plants. De Candolle held that the starting-point should be the temperature of germination in each plant. Thus, in the case of barley, it would be 5°; in that of wheat, 6°, and all temperatures below this critical point[1] should be disregarded in obtaining the accumulated temperatures.

The question as to what critical temperature should be assumed for given plants has been made the subject of an ingenious investigation by Arthur von Oettingen. If the accumulated temperatures are calculated on the basis of different probable critical temperatures, that value will prove to be the most constant, *i.e.*, will show the minimum mean departure from year to year, which has been obtained by starting with the true critical temperature. Thus, for example, the accumulated temperatures for the first blossom of *prunus padus* showed the following mean departures in seven years:—Critical temperature, 0°; accumulated temperature, 334°; mean departure, ±13°. Critical temperature, 2°; accumulated temperature, 234°; mean departure, ±5°. Critical temperature, 4°; accumulated temperature, 162°; mean departure, ±6°. Critical

[1] The word "temperaturschwelle" cannot be translated literally.

temperature, 6°; accumulated temperature, 111°; mean departure, ± 7°. In short, the accumulated temperature which was based on the critical temperature of 2° proves to be the most constant in the different years, and therefore gives the best values for use in comparisons. Further, the probable error in the date of the beginning of blossoming is the least for this same critical temperature. The method advocated by von Oettingen seems, on account of its many details, not to have been subjected to careful proof by others. In general, there is a tendency in recent years not to attribute so much importance to the phenological accumulated temperatures as was once done. Nevertheless, when they are subjected to adequate criticism, and when regard is had to their conditional meaning, these accumulated temperatures should not be wholly discarded if we wish to indicate the heat which different plants, especially the cultivated varieties, need. As the matter cannot be further discussed here, the reader is referred to the following recent publications on phenology :—

Bibliography.—S. Günther : *Die Phänologie*, Münster, 1895—a short, concise account of researches in this field. O. Drude : *Handbuch der Pflanzengeographie*, Stuttgart, 1890, pp. 17-48 ; and *Deutschland's Pflanzengeographie*, Part I., Stuttgart, 1896 ; Section V., *Die Periodische Entwickelung des Pflanzenlebens im Anschluss an das mitteleuropäische Klima.* A. von Oettingen : *Phänologie der Dorpater Lignosen. Ein Beitrag zur Kritik phänologischer Beobachtungs- und Berechnungsmethoden*, Dorpat, 1879. H. Hoffmann : *Phänologische Untersuchungen*, Giessen, 1887. R. Hult: "Récherches sur les phénomènes périodiques des Plantes," *Nova Acta R. Soc. Sc. Upsalae*, Series III., 1881. E. Ihne : "Phänologische Jahreszeiten," *Potonié Naturw. Wochenschrift*, X., No. 4, 1895; also, "Karte der Aufblühzeit von *Syringa vulgaris* in Europa," *Bot. Zentralblatt*, 1885, Vol. XXI., 3-5. H. Hoffmann ; "Vergleichende phänologische Karte von Mitteleuropa," *Peterm. Mitt.*, XXVII., 1881, 19-26. M. Staub : "Phänologische Karte von Ungarn," *ibid.*, XXVIII., 1882, 335-339. Very instructive phenological charts were published by A. Angot in his paper entitled "Résumé des Études sur la Marche des Phénomènes de Végétation et de la Migration des Oiseaux pendant les dix Années 1881-90," in the *Ann. Bur. Centr. met. de France*, 1892, I. *Mémoires*, Paris, 1894, B159-B210; also, Angot's great work "Étude sur les Vendanges en France," *ibid.*, 1883, I. *Mémoires*, Paris, 1885, B29-B120, which is important in the study of changes in climate as well. The numerous works of Fritsch, Linsser, etc., bearing earlier dates, cannot be referred to here. The Annual Reports on Phenological Observations in the British Isles by E. Mawley are published in the Quarterly Journal of the Royal Meteorological Society for recent years. See also C. Abbe, in *Maryland Weather Service*, Vol. I., 1899, 267-278. Professor Abbe has prepared an unpublished report of great value (officially known as No. 5119, Sig. 91) on "The Relations between Climates and Crops," dealing with the physiological and experimental work which has been carried on in laboratories, and also with the results of experience in the open air under natural climatic conditions.

PART II.

GENERAL CLIMATOLOGY.

SOLAR CLIMATE AND THE CHIEF VARIETIES OF PHYSICAL CLIMATE.

SECTION I.—SOLAR CLIMATE.

CHAPTER VI.

SOLAR OR MATHEMATICAL CLIMATE.

Definition of solar climate.—If the surface of the earth were occupied altogether by land, and if there were no surrounding atmosphere, the condition of our planet would be somewhat similar to that of the moon at the present time. Under these conditions, the distribution of temperature over the earth would depend solely upon the amount of heat received from the sun at any given place, and upon the loss of heat by radiation at that place. As these two factors would necessarily be the same at all points along the same parallel of latitude, the zones of equal temperature would coincide with the parallels of latitude. Even the presence of a vapourless atmosphere would interfere but little with this distribution of temperature, for only the absolute amounts of heat received at, and radiated from, the surface of the earth would thereby be affected. It is true that convectional currents would be produced under these conditions; but as there would be no reason for the more frequent occurrence of warm or cold air currents along some meridians than others, the distribution of temperature in zones bounded by the parallels of latitude would not thereby be interfered with.

In so far as climate depends only upon the amount of solar radiation received at any place by reason of its latitude, it is called *solar climate*. Under the hypothetical conditions above described, therefore, solar climate alone would prevail upon the earth. The amount of solar radiation which would be received at any place during a day, or a year, if there were no atmosphere, can readily be calculated. Further, the amount of solar radiation which would be transmitted through a per-

petually cloudless and vapourless atmosphere can also be approximately deduced from the foregoing data. As a matter of fact, however, the direct radiation from the sun which falls upon the earth's surface is decidedly affected by the widely varying degrees of clearness and vapour contents of the atmosphere, both in time and place, *i.e.*, by the differences in its transmissive quality. Furthermore, the temperature of the air at any given place does not depend solely upon the amount of absorption and of radiation at that place, but also upon the interchange of heat which goes on by means of air and ocean currents between places of different temperatures. It results from this that, as is well known, the temperature of the air is no longer the same everywhere along the same parallel of latitude, especially not in higher latitudes, and that even the amount of direct solar radiation at the earth's surface is no longer the same along the same parallel. Nevertheless, solar climate must furnish the basis for the description of the existing distribution of temperature on the earth's surface, for all the phenomena upon the earth which are connected with temperature depend upon the solar rays, and the simplest laws of the distribution of solar radiation are likewise the laws which chiefly control the distribution of temperature. Not only so, but the value of insolation on clear days, *i.e.*, its maximum value, is determined for every place by the solar climate of that place, for the state of the sky may lessen the amount of radiant energy received, but cannot increase it.[1]

Ptolemy's climatic zones.—Ptolemy and the ancient geographers classified their climatic zones solely with reference to the conditions of illumination which these several zones receive. The word *climate* (from *κλίνειν, to incline*) still recalls the fact that it was meant to describe only the differences in insolation which are due to the inclination of the earth's axis. The climates of Ptolemy were zones in which the length of the longest day increased successively by half an hour between the equator and the Arctic circle. The zones were therefore of very different widths. The first climate embraced $8\frac{1}{2}$ degrees of latitude; the fifteenth, but one degree (from 61° to 62°), and the twenty-fourth, but three minutes.

The laws governing the changes in the intensity and amount of insolation with changes in latitude have been investigated by Halley, Lambert, Meech, and Wiener.[2]

[1] This statement needs a slight qualification, which will be noted later.

[2] Meech: "On the Relative Intensity of the Heat and Light of the Sun," *Smithsonian Contributions*, Vol. IX., Washington, 1857; Chr. Wiener: *Ueber die Stärke der Bestrahlung der Erde durch die Sonne in ihren verschiedenen Breiten*

We shall here refer only to the most important results of these mathematical investigations.

Distribution of insolation over the earth.—The intensity and amount of insolation received by any portion of the earth's surface depends upon the angle of incidence of the sun's rays, or the sun's altitude, and upon the duration of the insolation, or the length of the day. The law by which the intensity of light and of heat depend upon the angle of incidence of the sun's rays is known through the teachings of physics, and is illustrated in the accompanying figure.

The surface BB receives less insolation in proportion as this surface is larger than the surface AA, at right angles to the same pencil of rays, S. The intensity of insolation, I', on the surface B, is to the intensity I, on the surface A, inversely as the areas of these surfaces, or

$$I' : I = A : B,$$

or $$I' = I(A \div B) = I \sin h,$$

when the rays make the angle h with the horizontal.

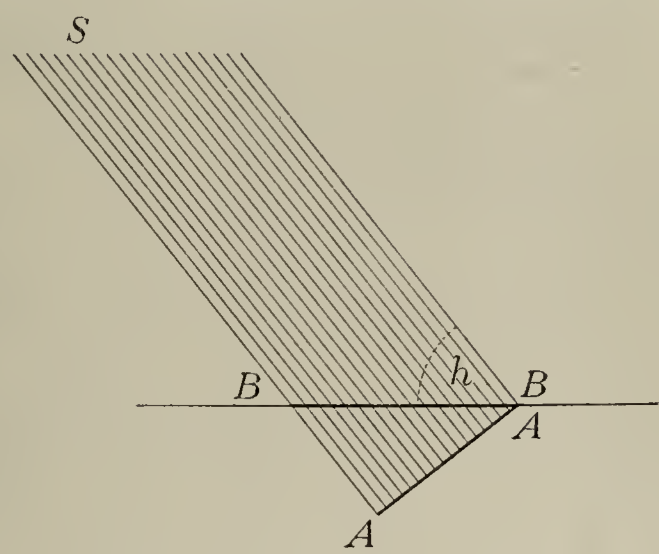

Fig. 1.

The intensity of insolation, therefore, varies with the sine of the sun's altitude. The greatest altitude which the sun can reach at noon decreases with the latitude, being equal to the meridian altitude of the equator $+ 23\frac{1}{2}°$, or to $90° -$ the latitude $+ 23\frac{1}{2}°$. If the sun were fixed in the equator, and if day and night were equal everywhere on the earth's surface, as is actually the case at the equinoxes, insolation would vary with latitude according to a very simple law, namely, directly as the sine of the meridian altitude of the sun. When the sun is on the equator, this altitude is equal to the meridian altitude of the equator, or $90° -$ the latitude. As this condition would obtain on every day in the year, the annual amount of insolation would vary with the sine of the meridian altitude of the equator, or with the cosine of the latitude. Since the sun does not move far from the equator, and the stronger insolation of summer is partially compensated by the weaker insolation

und Jahreszeiten, Karlsruhe, Bielefeld, 1876; also, *Z.f.M.*, XIV., 1879, 113-130; W. Zenker: *Der Thermische Aufbau der Klimate aus den Wärmewirkungen der Sonnenstrahlung und des Erdinnern*, fol., Halle, 1895, pp. 252, Charts 5, [*M.Z.*, XIII., 1896, (25)-(28)].

of winter, this law of the distribution of temperature, which really applies only to the period of the equinoxes, holds fairly well for the distribution of the annual amounts of insolation as far as latitude 25°, if we assume an error of one per cent.

That this simple law cannot hold in higher latitudes is apparent when we remember that, if it did, the poles would receive no insolation at all. As a matter of fact, the poles receive a very considerable amount of insolation during their long summer days, although this

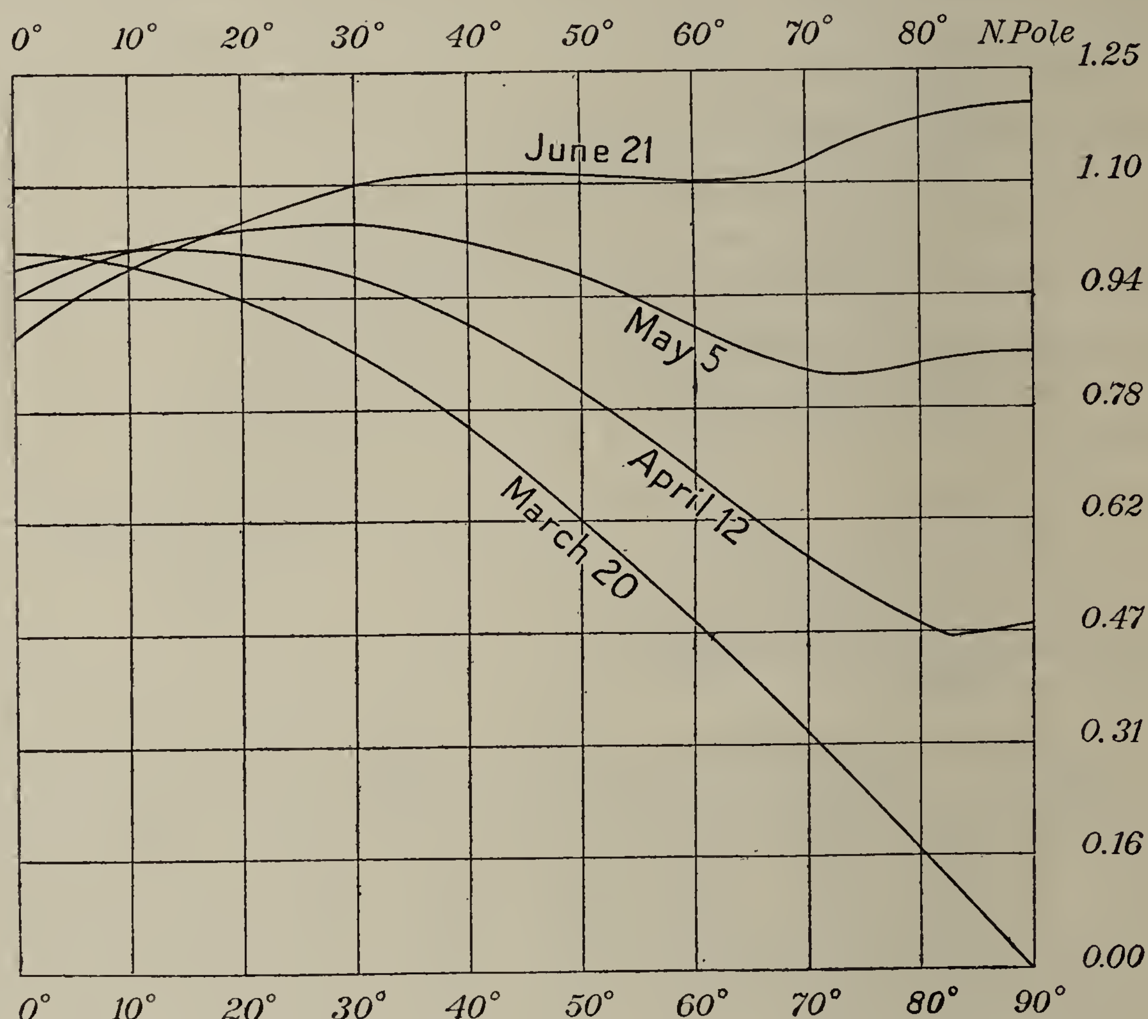

Fig. 2.

condition lasts for a short time only. The law by which the daily amounts of light and of heat are distributed over the earth's surface when the sun is not on the equator is a very complex one, for the reason that the rapidly increasing length of day toward one of the poles soon more than compensates for the decreasing angle at which the solar rays strike the earth. Thus there come to be two maxima of the total daily insolation, one in lower latitudes, and the other at the pole. The accompanying figure (after Wiener) shows the distribution of the daily amounts of insolation received from the equator to the north pole, on March 20, April 12, May 5, and June 21. The relative

intensities of the insolation are entered at the right-hand margin of the figure.

It appears from this figure that on June 21 the maxima of insolation are at latitude 43° 30′ and at the pole. There is a slight minimum of insolation at latitude 62°, but still the insolation at this point is considerably greater than the amount at the equator, at the same time. From latitude 43° 30′ N. the amount of insolation decreases continuously till it reaches zero on the Antarctic Circle. If we call the total daily insolation received at the equator on March 20, 1000, the relative distribution of the daily amounts of insolation at the most critical latitudes on June 21 will be expressed by the following figures:—[1]

North Pole	62°	$43\frac{1}{2}$°	Equator	$66\frac{1}{5}$° S.
1203	1092	1109	881	0

The insolation at the north pole on June 21 is therefore more than 20 per cent. in excess of the largest amount which the equator ever receives, and is 36 per cent. in excess of that received at the equator on June 21. Furthermore, during 28 days before and after the turn of the sun, *i.e.*, during 56 days, the insolation at the pole which is having its summer is stronger than that at any other place on the earth's surface, and during 84 days it is greater than the amount at the equator at the same time.' For the north pole, this latter period lasts from May 10 to August 3.

The intensity of insolation during the summer of the southern hemisphere is somewhat greater than that in the northern. In winter, the conditions are reversed. This results from the fact that the earth is nearer the sun during the summer of the southern hemisphere, while it is farther from the sun during the summer of the northern hemisphere. The difference in the intensity of solar radiation, at the extreme points, namely on January 1

[1] If we choose as our unit the amount of heat needed to raise the temperature of one gram of water 1°, the above figures will also represent the daily amounts of heat received on a surface of one square centimeter in area, with the possibility of error due to the uncertainty which still exists as to the absolute value of solar radiation. The figures could therefore be regarded as absolute units of heat, and would then correspond to a solar constant of 2·16 units of heat per square centimeter a minute. Langley's value of about 3 calories for the solar constant, which was determined by means of very ingenious observations, continued for many years, is now generally accepted as the most likely one. If Langley's value be used, the foregoing figures must be increased in the proportion of $3 : 2{\cdot}16 = 1{\cdot}39$, or, roughly, 40 per cent.

(perihelion), and on July 3 (aphelion), amounts to about 1/15 of the total insolation.[1]

This difference is sufficiently large to be noticeable. According to Dove, the marked change in the temperature which is felt in stepping from the shade into the sunlight during the summer of the southern hemisphere surprises immigrants into Australia and New Zealand. The heating of the soil, and the temperature maxima, are greater in Australia and South Africa than in corresponding latitudes in the northern hemisphere, notwithstanding the fact that the mean temperature of summer in the southern hemisphere is, for other reasons, notably lower.

The intensities of insolation at different points on the earth's orbit are inversely proportional to the squares of the sun's distances, or, directly proportional to the squares of the sun's apparent radius. If, for example, we wish to know the amount of insolation at the south pole on December 21, the result is reached by means of a simple calculation depending upon the sun's apparent radius, which can be obtained from any astronomical almanac. The radius of the sun's disc subtends an angle of 946″ on June 21, and an angle of 978″ on December 21. We have, therefore, as an expression for the insolation at the south pole on December 21,

$$1203\left(\frac{978}{946}\right)^2 = 1286.$$

Thus the south pole receives at this time 83 more radiation units than the north pole receives on June 21. As a result of the changes in the distance between the earth and the sun, the annual march of insolation becomes unsymmetrical in the two halves of the year; in low latitudes, where the annual variation is slight, the whole form of the annual curve is thereby very decidedly modified.

The following table gives the amounts of insolation received on

[1] If the eccentricity be designated by ϵ, the distance when the sun is nearest is $1-\epsilon$; and when the sun is farthest, $1+\epsilon$. The intensities are inversely proportional to the squares of the distances; hence the intensity at perihelion is $\left(\frac{1+\epsilon}{1-\epsilon}\right)^2$ if the intensity at aphelion is 1. The result is, approximately, $1+4\epsilon$. As ϵ is at present about 1/60, the fraction given in the text is obtained. If, following Laplace, we take as the maximum of the eccentricity 0·07775, *i.e.*, about 1/13, the intensity of insolation at perihelion would equal aphelion $(1+4/13)$. Thus the warming, when the earth is near the sun, would be almost 1/3 greater than at aphelion, and the differences between the extremes would become very great.

the equator, and at latitudes 15° and 20° north and south, at the solstices, and when the sun is overhead.

AMOUNTS OF INSOLATION AT DIFFERENT LATITUDES AND ON DIFFERENT DAYS.

Latitude.	Insolation, Dec. 21.	Culmination.		Insolation, June 21.	Culmination.	
		Date.	Insolation.		Date.	Insolation.
20° N.	677	May 20	1041	1045	July 23	1034
15° N.	749	May 1	1017	1012	Aug. 12	1008
Equator	942	March 20	1000	881	Sept. 23	988
15° S.	1081	Feb. 8	1061	701	Nov. 3	1053
20° S.	1116	Jan. 20	1101	633	Nov. 21	1094

At the equator there is a double period of insolation, two maxima at the equinoxes and two minima at the solstices. The differences between these two extremes result from the variations in the distance of the sun at different times of the year. These variations are also sufficiently important perceptibly to change the dates upon which the maxima and the minima of insolation occur. The first maximum comes somewhat before March 20; the second does not come until October 14, instead of on September 23, which is the date of the sun's culmination. At the equator, the maximum difference between the amounts of total daily insolation received during the year is only 12 per cent. In the vicinity of the equator also, roughly as far as latitude 12° north and south, there is a double annual period; but at latitude 15° this period is single, as is shown by the figures above. The intensity of insolation in these latitudes is nearly constant throughout the whole summer. At latitude 20°, the difference between the extreme values of the daily insolation reaches 368 and 483, which is three to four times as great as at the equator.

The following figures show how the differences between the amounts of insolation at the summer and the winter solstices increase with the latitude outside of the tropics. These figures show the total daily insolation received at every 10° of latitude on June 21 and December 21.

AMOUNTS OF INSOLATION ON JUNE 21 AND DECEMBER 21 AT DIFFERENT LATITUDES IN NORTHERN AND SOUTHERN HEMISPHERES.

NORTHERN HEMISPHERE.

	30°	40°	50°	60°	70°	80°	90°
June 21, -	1088	1107	1105	1093	1130	1184	1202
Dec. 21, -	520	355	197	56	0	0	0
Difference, -	568	752	908	1037	1130	1184	1202

SOUTHERN HEMISPHERE.

	30°	40°	50°	60°	70°	80°	90°
Dec. 21, -	1163	1183	1180	1168	1207	1265	1284
June 21, -	487	332	184	52	0	0	0
Difference, -	676	851	996	1116	1207	1265	1284

As radiation, which is the second factor upon which temperature depends, increases with the latitude, the actual differences between the temperatures of summer and of winter are much greater than would appear from the differences above noted. This table likewise shows that, so far as solar climate is concerned, the contrasts between the insolation of summer and of winter are about 7 to 17 per cent. greater in the southern hemisphere than in the northern. Furthermore, we must take into account the longer continuance of radiation during the southern winter, which is about eight days longer than the northern winter. For these reasons, the solar climate of the southern hemisphere is more extreme than that of the northern.

The annual march of the air temperature on the earth's surface is, first of all, dependent upon the heat of the sun. It is, however, also controlled by many secondary meteorological phenomena, which prevent it from being precisely like the annual march of the insolation. Secondary phenomena, such as the beginning of the rainy season, for example, may, to a marked extent, disturb the annual march of temperature as determined directly by the sun. This is especially the case in low latitudes, where the variations in temperature in the course of a year are slight. In the higher latitudes, the extremes of air temperature coincide more nearly with those of insolation; yet the former always lag behind the latter. This lagging amounts to about one month on the average.

Annual amounts of insolation.—The direct computation of the total yearly insolation at different latitudes is much more difficult than the computation of the daily insolation, for the reason that no simple formula can be derived which will lead directly to the determination of these annual amounts by any convenient method.[1] We can, however, by an indirect method, and without the use of elaborate mathematical calculations, gain some knowledge of these annual amounts of insolation, or of the amounts during single parts of the year, in the following way. The intensities of insolation for half-months, or better still, for ten-day periods, are determined by means of the formula given in the Appendix to this Chapter (see p. 124). The annual variation of insolation is then represented graphically by entering the amounts thus determined successively at equal intervals as ordinates above a horizontal axis of abscissae. If the ends of these ordinates are joined by a free hand curve, a diagram is constructed whose area is proportional to the total yearly insolation. The total yearly insolation may be obtained as follows: Measure, by means of a planimeter, the surface area of this figure, and divide by the area of the rectangle, one of whose sides is the length of the year, according to the scale which has been adopted, and the other of which (the ordinate) is the unit of insolation. By an analogous method, the same figure may be used to give the amounts of insolation for the months and for the seasons. The annual curve, as well as the rectangle above mentioned, may also be drawn, on a larger scale, upon heavy paper of uniform thickness, and then cut out. The surfaces may

[1] An approximate formula by means of which the annual amounts of insolation (W) may be calculated for every degree of latitude as far as the Arctic and Antarctic Circles is here given. It is the formula of Haughton, with some rearrangement of terms and some abbreviation.

$$W = C \cos \phi \, (1 + 0{\cdot}04366 \tan^2\phi + 0{\cdot}00049 \tan^4\phi).$$

C is the annual amount of insolation received at the equator, *i.e.*, 160,582 A; in which A is the solar constant (see Appendix) and ϕ is the latitude. The formula does not hold for latitudes at which the sun can remain above the horizon for 24 hours. For the latitude of Vienna, for example, we obtain ($\phi = 48° \, 12'$),

$$W = 107{,}034 \text{ A } (1 + 0{\cdot}0436 \tan^2\phi + 0{\cdot}00049 \tan^4\phi),$$

which gives 112,960 A. If the sun were fixed in the equator, and if therefore the amount of insolation decreased simply as the cosine of the latitude, the annual amount at the latitude of Vienna would be only 107,034 A calories. The obliquity of the ecliptic increases this total amount to the extent of 5,926 A calories, or 5½ per cent. If we adopt Langley's value, A = 3, the annual amount for Vienna becomes 338,880 calories.

then be weighed. These weights are proportional to the total yearly insolation.

The following table gives the total yearly insolation for every five degrees of latitude, expressed in thermal days (Meech). The unit is the average total daily insolation at the equator.

ANNUAL AMOUNTS OF INSOLATION.

EQUATOR = 365·24.

Latitude.	Thermal Days.	Difference.	Latitude.	Thermal Days.	Difference.
5°	364·0	1·2	50°	249·7	20·1
10°	360·2	3·8	55°	228·8	20·9
15°	353·9	6·3	60°	207·8	21·0
20°	345·2	8·7	65°	187·9	19·9
25°	334·2	11·0	70°	173·0	14·9
30°	321·0	13·2	75°	163·2	9·8
35°	305·7	15·3	80°	156·6	6·6
40°	288·5	17·2	85°	152·8	3·8
45°	269·8	18·7	90°	151·6	1·2

The pole is thus seen to receive $41\frac{1}{2}$ per cent. of the measure of insolation which reaches the equator, while it would have no insolation at all if the sun always remained on the equator. The greater the obliquity of the ecliptic, the greater is the amount of insolation received at the pole. When the obliquity of the ecliptic reaches its maximum value of 24° 50′ 30″, the pole has 160·0 thermal days, or 8·4 more than it has now. The equator then has 1·7 less, while about latitude 40° there is no change.

The above table further shows that there is but little variation in the annual amounts of insolation with latitude near the equator and near the pole. The most rapid variation is between latitudes 50° and 60°.

Annual amounts of insolation in northern and southern hemispheres.—Both hemispheres receive equal amounts of insolation at the same latitude, notwithstanding the difference in the intensity of insolation during the same seasons. While the sun is south, during the summer of the southern hemisphere, the insolation is truly more intense; but the sun remains south of the equator nearly 8 days less than it does north of the equator, because the earth moves more rapidly around its orbit when it is nearer the sun. We learn from astronomy that the angular velocity of movement of the earth around its orbit varies

inversely as the square of the earth's distance from the sun, *i.e.*, in precisely the same ratio as the intensity of the insolation. The amount of heat received during any given period of time varies at every point of the orbit in precisely the same ratio as that by which the earth's longitude increases during that time. Thus, equal angles described by the radius vector always correspond to equal amounts of insolation. If a line be drawn from any point on the earth's orbit through the sun to the opposite point, the orbit will be divided into two portions which receive equal amounts of insolation. The inequality in the intensity of insolation over the two portions is exactly counterbalanced by the compensating inequality in the length of time which the earth takes to pass over these two portions. Thus it happens that the northern hemisphere, in its summer half-year, receives the same total insolation as the southern hemisphere in its summer half-year. This likewise holds true for the winter half-years, and for the astronomical and the meteorological quarters.

The mean intensity of insolation for the whole torrid zone as far as the tropics, is, according to Meech, 356·2 thermal days; that for the temperate zone is 276·4; and that for the frigid zone, 166·0 days. The average for the whole earth is 299 days, or $\frac{5}{6}$ths of the insolation at the equator.

Measurement of the intensity of solar radiation: Solar constant.[1]—Among the results which the efforts to measure the absolute intensity of solar radiation have thus far given, those obtained by Langley seem, at present, to be the most trustworthy. Langley found that the value of solar radiation at the upper limit of the atmosphere is 3 units of heat per square centimeter per minute.[2]

The cross-section of the pencil of rays which reaches the earth has the same surface area as a great circle of the earth, *i.e.*, one-quarter of the earth's surface. The sum-total of heat developed by radiation on the earth in one year is therefore:

$$3 \times 60 \times 24 \times 365{\cdot}2 \times R^2\pi = 2011 \times 10^{21} \text{ calories,}$$

[1] A summary of the older attempts to measure the intensity of solar radiation may be found in Schmid's *Lehrbuch der Meteorologie*, 123-134. An excellent account of similar investigations to the year 1875 is given by Radau in his *Actinométrie* and his *Les Radiations chimiques du Soleil*, Paris, 1877. A comprehensive summary of the whole matter is contained in O. Chwolson's "Ueber den gegenwärtigen Zustand der Aktinometrie," *Repert. f. Met.*, XV., No. 1, St. Petersburg, 1892. See also Violle's Report on Solar Radiation, presented at the International Congress of Meteorology, Paris, 1900.

[2] S. P. Langley: "Researches on Solar Heat and its Absorption by the Earth's Atmosphere," Washington, D.C., 1884. Professional Papers of the Signal Service, No. XV. (*M.Z.*, III., 1886, 193-207).

or more than two quadrillion units of heat, reckoned in cubic centimeters and centigrade degrees. This amount of heat would suffice to melt a layer of ice 5380 cm. thick over the whole earth. The equator receives annually 481,750 calories, and this heat would melt a layer of ice around the equator 66 m. in thickness.[1]

[1] Trabert has recently answered the question concerning the distribution of temperature, as dependent upon the sun, over an earth not possessing an atmosphere.

The quantity of heat developed at any latitude per day or per month is known as soon as a certain solar constant, *e.g.*, Langley's value of 3, is adopted. The actual temperature at any point on the earth's surface, however, is the resultant of this heat received from the sun, and the loss of heat by radiation. The temperature rises under the influence of insolation until the loss of heat by radiation just balances the amount of heat received. If, therefore, we assume a given law of radiation, preferably that of Stefan, who has found radiation proportional to the 4th power of the absolute temperature, T, and if we express the amount of heat received by Q, in gram-calories, we have the equation,

$$Q = aT^4.$$

In this equation, a is a coefficient which Stefan has found to be 723 in units of the 13th decimal place, *i.e.*, 723×10^{-13}. From this we can obtain T, *i.e.*, $273° + t$, according to the usual scale of temperature, when we make a proper substitution for Q, *e.g.*, the quantities of heat computed by Angot.

By this method the mean temperature of the whole earth without an atmosphere is found to be 46°; that of the day-half, 178°; and that of the night-half, -86°. Incidentally, these figures give the mean diurnal range of temperature between day and night. Below are given the extreme values of the mean monthly temperatures for every 10° of latitude, and also the annual means. (For the actual computation the reader is referred to *M.Z.*, XI., 1894, 426.)

TEMPERATURE OF THE EXTREME MONTHS ON AN EARTH WITHOUT ATMOSPHERE.

Equator.	10°	20°	30°	40°	50°	60°	70°	80°	Pole.
67	67	70	74	75	75	73	76	80	82
56	50	36	16	-10	-45	-103	-273	-273	-273
ANNUAL MEANS.									
62	61	57	50	39	24	1	-43	-81	-105

According to Trabert, these temperatures represent the actual solar climate on an earth without an atmosphere. The effect of the atmosphere upon temperatures observed on the earth's surface is clearly shown by the results of these computations; but the length of the day is insufficient to establish a fixed subsurface temperature gradient, and the maximum surface temperature given by theory, except in the polar zones.

The effect of the earth's atmosphere upon the amount and the character of solar radiation received at the earth's surface is twofold, quantitative and qualitative. If the sky be cloudless, the longer the path of the sun's rays through the atmosphere, the greater is the quantity of radiant energy absorbed by the atmosphere, as compared with the absorption by a smaller air mass of the same quality. The altitude of the sun is therefore of importance for two reasons, viz., because both the angle of incidence of the sun's rays, and the length of the path which these rays have to travel through the atmosphere, depend upon the sun's altitude. The first of these subjects has already been discussed. The intensity of solar radiation, therefore, decreases with decreasing altitude of the sun very much more rapidly than would be the case were there no atmosphere.

The following figures give, according to Zenker,[1] the relative thicknesses of the atmospheric strata through which the rays have to pass with varying altitudes of the sun, as well as the amount of the transmitted radiation.

RELATIVE DISTANCES TRAVERSED BY SOLAR RAYS THROUGH THE ATMOSPHERE AND INTENSITIES OF RADIATION PER UNIT AREAS.

Altitude of Sun.										
0°	5°	10°	20°	30°	40°	50°	60°	70°	80°	90°
Relative Lengths of Path of Rays through the Atmosphere.										
44·7	10·8	5·7	2·92	2·00	1·56	1·31	1·15	1·06	1·02	1·00
Intensity of Radiation on a Surface Normal to the Rays.										
0·0	0·15	0·31	0·51	0·62	0·68	0·72	0·75	0·76	0·77	0·78
Intensity of Radiation on a Horizontal Surface.										
0·0	0·01	0·05	0·17	0·31	0·44	0·55	0·65	0·72	0·76	0·78

The last line of figures is obtained by multiplying the line just above by the sine of the sun's altitude (or the cosine of the zenith distance). The values given are based on Langley's observations, and

[1] W. Zenker: *Die Vertheilung der Wärme auf der Erdoberfläche*, Springer, Berlin, 1888.

therefore assume a fairly clear atmosphere and a low degree of humidity. As a general rule, the average coefficient of transmission when the rays fall vertical is taken as only 0·75, or even 0·72. Very believes that with a moist atmosphere the coefficient of transmission is as small as ·50. It follows from the foregoing table that, as the result of the interposition of the atmosphere, the intensity of direct solar radiation depends upon the sun's altitude to a much greater extent than was formerly assumed.[1] Therefore latitude is now seen to play a more important part, and the higher latitudes are evidently more unfavourably situated, than would appear from the older tables of the intensities of solar radiation, except at those seasons when they enjoy an exceptionally transmissive atmosphere.

The variations in the intensity of light and heat with the variations in latitude, already noted, therefore suffer considerable modifications at the earth's surface. Assuming 0·7 as the coefficient of transmission, we obtain, for example, the following table (after Angot):—

VALUES OF TOTAL DAILY SOLAR RADIATION AT THE UPPER LIMIT OF EARTH'S ATMOSPHERE AND AT SEA LEVEL.

Latitude, - -	Equator.	40°	N. Pole.	Equator.	40°	N. Pole.
	Upper Limit of Atmosphere.			Earth's Surface.		
Winter Solstice, -	948	360	0	552	124	0
Equinoxes, - -	1000	773	0	612	411	0
Summer Solstice, -	888	1115	1210	517	660	494

[1] The dependence of the intensity of solar radiation upon the thickness of the atmospheric strata through which the sun's rays have to pass, is, in general, expressed by the following equation, in which q is the coefficient of transmission, d, the thickness of the atmospheric strata, and I, the intensity of insolation at the upper limit of the atmosphere:

$$I' = Iq^{d} = Iq^{\sec z}.$$

If the zenith distance of the sun is expressed by z, then $d = \sec z$ is approximately true. If every stratum transmits the qth portion of the solar radiation, the second stratum receives only qI. This second stratum lets only the qth part through, therefore the third stratum receives only qqI, *i.e.*, q^2I; the fourth, q^3I, and only the Iq^4th part goes through the fourth stratum. If we imagine the whole atmosphere divided into n such strata, n also represents the thickness of the atmosphere, and we have as an expression for the transmitted solar radiation Iq^n or Iq^d, as given above. This simple formula, however, takes no account of the actual complexity of the absorbent process, and especially of the complete disappearance of certain regions of the infra-red spectrum through the selective

It is a difficult matter to make a direct calculation of the total amount of radiant energy absorbed as heat between sunrise and sunset at different latitudes. Pouillet estimated this amount at 0·5 to 0·4 of the total radiation; but Angot was the first to complete the laborious computations.

The following observations, made by Crova, at Montpellier, show that Pouillet's estimate is correct if we limit the radiation to the remnant which has escaped band absorption, and their results, as determined by the present author, are the basis of the data given above for the earth's surface. By means of hourly observations on two perfectly clear days at opposite seasons of the year, Crova determined, with the greatest possible accuracy, the amount of heat which was received upon a horizontal surface during these days. For purposes of comparison, we have determined the total quantity of heat received from the sun at the latitude of Montpellier (43° 36′) on these two days, using for the solar constant the value 2·24, which was obtained as the result of Crova's own observations, but which must be taken as an estimate of the above-mentioned remnant. From these data, the quantity of heat which was absorbed by the atmosphere during the whole day in both these cases can at once be seen.

QUANTITY OF SOLAR RADIATION TRANSMITTED BY THE ATMOSPHERE AT MONTPELLIER.

	January 4.	July 11.	
Solar radiation observed,	161·2	574·1	calories.
,, ,, calculated,	322·3	1122·0	,,
Transmission ratio,	0·50	0·51	,,

Notwithstanding the great difference in the length of the day (8 h. 52 m. and 15 h. 2 m.) and in the sun's altitude, almost exactly one-half of the radiation from the sun was in both cases absorbed by the atmosphere in the course of the day. The amount of heat reaching a surface kept always normal to the rays, was 535·0 and 876·4 on the two days, according to Crova's observations. The quantities of heat, calculated on the basis of the length of the day, are 1191·6 and 2020·5. The corresponding ratios are therefore 0·45 and 0·43. It is thus seen that in the case of a body normal

absorption by aqueous vapour, as will be shown more fully in discussing the qualitative effect of atmospheric absorption.

to the sun's rays, nearly 0·6 of the apparent solar radiation is lost, and to this must be added the loss in the upper air which is not taken into account in these methods of reduction.

It may be said that, as a general rule, in middle latitudes, even when the sky is perfectly clear, the atmosphere absorbs considerably over one-half of the energy radiated daily to the earth from the sun. We have tried to determine the relation between the daily radiation at the upper limit of the atmosphere and at the earth's surface, at the equator on the equinoxes. This was done by using 0·75 as the coefficient of apparent absorption, and computing the amounts of solar radiation which reach the earth's surface with varying altitudes of the sun. These amounts were then represented graphically, and the surface area of the figure was determined by means of a planimeter. The result gave 0·57, which means that even at the equator on the days when the sun passes through the zenith, only 57 per cent. of the remnant of insolation left after band absorption reaches the earth's surface.

Angot has advanced the investigations of Meech and Wiener by means of some very painstaking calculations. He has determined the amounts of heat received at different latitudes with reference to the absorption of solar radiation by the atmosphere. The results relate to the maximum and the minimum amounts received daily at twelve different times during the year, as well as to the total amount received during the summer and the winter half-years.[1] The values obtained by Angot for the intensity of solar radiation therefore correspond to the real solar climate on the earth's surface, as far as this can be determined by computations which necessarily omit all reference to the variations in cloudiness and in the quality of the atmospheric absorption; they tell us the quantities of heat which every tenth degree of latitude on the earth's surface receives directly from the sun on a perfectly clear day. Angot computed these values for various coefficients of transmission, namely, for $q = 0·9$, 0·8, 0·7 and 0·6. The different values thus obtained give us an approximate idea of the effect which varying degrees of atmospheric transparency (as regards the passage of radiation from the sun) have upon the quantity, as well as upon the distribution, of the intensity of insolation at the earth's surface, although a complete discussion of atmospheric action on radiation, in all of its complexity, would modify the

[1] Alfred Angot: "Recherches théorétiques sur la Distribution de la Chaleur à la Surface du Globe," *Ann. Bur. central mét. de France*, Tome I., 1883, Paris, 1885, B121-B169 (*M.Z.*, III., 1886, 540-546).

figures somewhat. On account of its importance from this point of view, an extract from the extended tables given by Angot is included in the Appendix to this chapter (p. 124).

In consequence of the absorption of solar radiation by the atmosphere, the decrease in intensity on the earth's surface is much more rapid towards the pole than we found it to be at the upper limit of the atmosphere. With a coefficient of transmission below 0·8, which represents the actual conditions, there is no longer a maximum at the pole on June 21, but insolation decreases continuously from latitude 40° to the pole, and is less there than at the equator at the same time. Exception must be taken to this general statement during the spring months, when the transmissive quality of the atmosphere improves with increase of latitude, owing to the small quantity of water vapour in the air of the polar regions at this season. The maximum of solar radiation on a normal surface occurs in the spring, in spite of the larger air mass then traversed by the rays. The relative amounts of heat received during a year at equator and pole, assuming different coefficients of transmission, are as follows[1]:—

QUANTITIES OF HEAT RECEIVED AT EQUATOR AND POLES WITH DIFFERENT COEFFICIENTS OF TRANSMISSION.

q.	1	0·9	0·8	0·7	0·6
Equator, -	350·3	298·4	251·9	209·2	170·2
Pole, - -	145·4	100·1	68·2	45·0	28·4

Taking as the coefficient of apparent transmission the value 0·7, which fairly well corresponds to the present condition of the atmosphere, the pole receives less than one-quarter of the heat received at the equator; and when $q = 0·6$, the pole receives but one-sixth of that amount. The less clear the atmosphere, the greater is the difference in temperature between poles and equator, in solar climate, and the more rapid is the decrease of temperature with latitude.

Let us now consider the conditions at some intermediate latitude, for example, at latitude 50°.

[1] The unit is the quantity of heat developed by radiation per equatorial day with the earth at an average distance from the sun, and with a declination of 0°. This unit is therefore different from that adopted by Meech, which was referred to on page 100. The quantity of heat received by the equator under these conditions is 458·4 × I, where I is the solar constant. If the latter is taken as 3, the unit adopted by Angot becomes equal to 1375·2 gram-calories.

QUANTITIES OF HEAT RECEIVED AT LATITUDE 50° WITH DIFFERENT COEFFICIENTS OF TRANSMISSION.

q =	1	0·9	0·8	0·7	0·6
Summer half-year, -	175·5	144·7	118·5	95·6	75·2
Winter half-year, -	64·1	45·9	32·7	22·9	15·4
Year, - - -	239·6	190·6	151·2	118·5	90·6

The ratio of winter insolation to summer insolation becomes a less and less favourable one, the less clear the atmosphere. If there were no atmosphere, the ratio of summer to winter radiation received at latitude 50° north would be 2·8. As a matter of fact, however, when $q = 0·7$ the ratio is 4·2; and when $q = 0·6$, the quantity of solar radiation received in summer is not much greater than that in winter without an atmosphere. The sum total for the year is only about one-half of the theoretical amount.

The maximum and the minimum daily amounts of heat[1] derived from direct solar radiation at latitude 50°, at the upper limit of the atmosphere, are 1112 and 198 respectively, for the northern hemisphere. This gives a ratio of 5·6 : 1. When $q = 0·7$, the amounts at the earth's surface are 633 and 41; ratio, 15·4 : 1. For the southern hemisphere these values are 1188 and 185, which give a ratio of 6·4 at the atmospheric limit. When $q = 0·7$, the values are 676 and 38 at the surface, or in the ratio 17·8, which, as already noted, is still greater. The smaller the coefficient of transmission, the less favourable is the relation of winter to summer. This may readily be understood from the fact that the influence of the low altitude of the sun in winter becomes more and more effective. Nevertheless, there is some compensation in winter from the improved transmissive quality of the air. All the data above refer only to the direct heat of radiation from the sun on perfectly clear days. The diffuse radiation of the atmosphere is not here taken into account.

Solar climate of Montpellier and Kiev.—Montpellier and Kiev are the only places whose solar climate is approximately known, as the result of direct measurements of the intensity of solar radiation

[1] In this computation, the total daily heat of solar radiation at the equator at the equinox, and with the sun at a mean distance, is taken as 1000. If we wish to get our results in calories we must, assuming 3 as the value of the solar constant, multiply Angot's figures by 1·375, *i.e.*, they must be increased by 37·5 per cent. This applies only to the values at the limit of the atmosphere.

regularly carried on for a number of years. Since 1883, the intensity of solar radiation at Montpellier has been measured, about noon on every clear day, by Crova and Houdaille. The diurnal variation of the intensity was determined, first, by hourly observations, and later, by self-recording instruments. As, furthermore, records of the duration of bright sunshine at Montpellier are also available, all the elements are at hand for the determination of the actual amount of solar radiation which reaches the earth's surface throughout the year at that place. On the basis of 11 years' observations, the mean intensity of solar radiation at noon has been found to be 1·08 calories; the maximum, 1·16, comes in April and May, and the minimum, 0·98, in December. The absolute maximum was 1·6 calories. The annual value of direct solar radiation, in calories, without considering the diffuse radiation, is 71,924. On an average July day, the mean value is 325; on an average December day, the mean value is but 61 gram-calories. Montpellier thus receives only about 20 per cent. of the solar radiation which theoretically belongs to its latitude (43° 36′). The average duration of sunshine is 50 per cent. of the possible duration. The coefficient of apparent transmission is greatest in winter (December, 0·71), and smallest in summer (0·48). In the mean for the year, the value is about 0·6.[1]

At Kiev, according to Savelief,[2] on the basis of three years' observations, the annual value of direct solar radiation is 60,745. On an average July day, the mean value is 328; and on an average December day, the mean is 13. If all the days were clear, Kiev would receive, on every sq. cm. of horizontal surface, 123·5 large calories from the sun. At the upper limit of the atmosphere, it receives 337·9. Thus it is seen that 63·5 per cent. of the total are absorbed, and only 36·5 per cent. reach the surface on wholly clear days. As a matter of fact, only 18 per cent. actually reach the surface.

Actinometrical observations at Pawlowsk and Katharinenburg.— Careful absolute measurements of the intensity of solar radiation have also been made by Schukewitsch at Pawlowsk, and Müller at Katharinenburg.[3] The direct intensity of radiation at Katharinenburg at noon attains a maximum of 1·44

[1] Cf. *M.Z.*, V., 1888, 198-199; *M.Z.*, IX., 1892, 203; *M.Z.*, XII., 1895, 184; *M.Z.*, XV., 1898, 72.

[2] *Comptes rendus des Obs. actinomet. faites en Kieff en* 1891 *et* 1892, St. Petersburg, 1893.

[3] *Repert. f. Met.*, XVII., No. 5, St. Petersburg, 1894. A discussion of the observations at Pawlowsk by Schukewitsch and Müller will be found in *Bull. St. Petersburg Akad.*, V. Ser., XI., No. 2, Sept., 1899.

calories at the end of March, and a minimum of 1·22 calories at the beginning of December. (At Pawlowsk, lat. 59° 41′ N., with the sun's altitude 30°, the values are 1·36 in April and a minimum of somewhat over 1·20 in July and August.) The intensity, when reduced to the same altitude and distance of the sun, shows a maximum at the beginning of January and a minimum in July. That the observed intensity at noon is less in summer, in spite of the greater altitude of the sun, than in March, results from the decrease in the diathermancy of the atmosphere in summer. The intensity is greatest in spring at Montpellier, as well as in Russia and in western Siberia. Occasionally, a secondary maximum occurs in the autumn.

Zenker's studies.—Zenker has recently computed the total radiation at the bottom of the atmosphere, for the year and for individual months, for every two latitude degrees from 70° to 10°. The table which was thus compiled is useful for many purposes.[1]

The attempt has been made to determine the climatic value of the heat of solar radiation, on the basis of these ascertained amounts of total solar radiation, and of the corresponding observed mean air temperatures. The most important points in this matter will be considered in a later paragraph.

Radiation from the atmosphere.—A considerable portion of the radiant energy which is withheld from the earth by the atmosphere, is replaced by the radiation from the atmosphere itself. The fine particles which are suspended in the atmosphere, such as minute drops of water, dust, etc.,[2] reflect and scatter the radiant energy from the sun. Thus the atmosphere itself becomes luminous, and is a source of light and of radiant heat. The great extent of the atmosphere makes this property one of considerable importance. The diffuse radiation from the sky is of special importance in the higher latitudes, where, by reason of the low altitude of the sun, the scattering and the absorption of direct solar radiation are very considerable. In the higher latitudes, moreover, the long duration of twilight to some extent compensates for this loss. If it were correct to estimate the energy of the diffuse radiation from the sky which is transformed into heat as equal to that which appears as

[1] W. Zenker: "Ueber den klimatischen Wärmewerth der Sonnenstrahlen und ueber die zum thermischen Aufbau der Klimate mitwirkenden Ursachen," *M.Z.*, IX., 1892, 336-344; 380-394. In Zenker's latest great work, *Der thermische Aufbau der Klimate*, the total monthly insolation is given for every 2° of latitude for the whole earth.

[2] According to Captain Abney it appears probable, as a result of his own observations at different altitudes above sea-level, as well as of laboratory experiments, that the minute particles which cause the selective scattering of light are water particles. These particles are probably very uniform in size, and very small as compared with the drops of water which form fog and cloud.

light, when the sun's altitude is $23\frac{1}{2}°$, the former would amount to 367 units of heat at the north pole on June 21. Under these conditions, the total radiation at the earth's surface would amount to 951 calories, as against 1202 at the upper limit of the atmosphere, a horizontal surface being assumed in both cases.

We are indebted to Maurer and Pernter for absolute measurements of atmospheric radiation at night. Maurer found the nocturnal radiation at Zürich, on a clear night in June, equal to 0·13 gram-calories per sq. cm. a minute. According to Stefan's law of radiation, the copper plate at a temperature of 15° would have had to lose by radiation a quantity of heat equal to 0·50 calories, if there had been no return radiation from the atmosphere. This return radiation amounted to 0·37 gram-calories ($0{\cdot}50 - 0{\cdot}13 = 0{\cdot}37$); *i.e.*, more than one-tenth of the solar radiation at the upper limit of the atmosphere, and this at night. Pernter's observations, at the summit of the Sonnblick (3100 m.), on a February night with a temperature of $-12°$, gave a radiation of 0·20 calories, and a return radiation from the atmosphere of 0·12 calories per sq. cm. a minute. Homén has recently made a study of the comparative values of the total radiation from sun and sky, and the radiation from the earth's surface.[1]

Even scattered clouds, or light clouds, are effective as reflectors of solar radiation, and hence the amount of cloud, as determined by the usual method of estimating cloudiness (*i.e.*, by its extent and not by its thickness), does not weaken insolation so much as is generally supposed. Indeed, it has actually been observed that, when the clouds are favourably distributed opposite the sun, the cloudiness may increase the intensity of radiation above the amount which would be possible if the sky were perfectly clear. As has been pointed out by Radau, the actinometrical measurements made at Montsouris show that the monthly means of estimated cloudiness are always larger than the calculated means based on observations with the actinometer. This is notably the case in the warmer half of the year. In summer, the cloudiness, as observed, is 5·6; and as calculated, it is 3·6. For the mean of the year, the results are: observed, 6·9; calculated, 5·0.

Relative values of direct and diffused sunlight.—According to Clausius, when the altitude of the sun is 40°, the diffused light amounts to about one-quarter of the direct sunlight upon a surface at right angles to the sun's rays. In the case of a horizontal surface, the ratio is two-fifths; and the smaller the coefficient of transmission, the larger is the ratio of diffuse light to direct sunlight.[2] The ratio of sunlight

[1] Th. Homén: *Der tägliche Wärmeumsatz im Boden und die Wärmestrahlung zwischen Himmel und Erde*, Leipzig, 1897.

[2] With 0·60 as the coefficient of transmission, and with the sun at an altitude of 30°, the direct illumination of a horizontal surface would be less than the illumina-

to the total skylight on a very clear October day, and upon a surface at right angles to the sun's rays, is given by Langley as one-fourth when the sun's altitude is 38°. The diffuse light increases the total illumination of a horizontal surface as much as if the sun were 5° higher above the horizon.

The chemical effects of diffuse light are still more important, for Bunsen and Roscoe found that the direct chemical effects of sunlight do not exceed those of the diffuse light until the sun's altitude becomes about 19°. In the case of an eye having the same relative sensitiveness for rays of different wave-length as that of the substances which are used to determine the chemical effects of radiation (*i.e.*, having the greatest sensitiveness for the blue and the violet), the illumination of a surface by direct sunshine would not equal the illumination by the light of the sky until the sun's altitude above the horizon became 19°.[1]

Measurements of the light of the sky.—Hardly any direct measurements of the luminous intensity of the sky and of direct solar radiation at different latitudes are available. The atmosphere, however, being very transparent to the luminous rays, the quantities of light computed for different latitudes by the formulas given by Meech serve as fair indications of the actual conditions of possible illumination in the different climates. This is the more true because of the fact that the diffused light partially compensates for the loss of direct insolation which results from the presence of the atmosphere. In higher latitudes, the light diffused by the atmosphere during the long twilight prolongs the duration of daylight considerably beyond the limits which it would have were there no atmosphere.

tion due to diffuse daylight. See also recent investigations by H. and O. Wiener: "Die Helligkeit des klaren Himmels, und die Beleuchtung durch Sonne, Himmel, und Rückstrahlung," *Abh. d. Kais. Leop.-Karol. Akad. d. Naturf.*, *Nova Acta*, LXXIII., No. 1, 4to, pp. 239, Leipzig, 1900.

[1] Recent observations by Brennand, in Dacca (India), gave the sun's altitude in this case as 13°. At lower altitudes, the skylight is chemically more active than direct sunlight. W. Brennand: "Photometric Observations of the Sun and Sky," *Proc. Roy. Soc.*, XLIX., 1891, 4-11 (*M.Z.*, VIII., 1891, 185-188). Some recent observations at Rome, Catania, and at the Etna Observatory are discussed by Q. Majorama, "On the Relative Luminous Intensities of Sun and Sky," *Philosoph. Mag.* (London), I., 6th Ser., 1901, 555-562 (translated from the Italian in *Rendiconti R. Accad. Lincei.*, *Cl. de Sci. fis.*, *mat. e nat.*, IX., 2nd Sem., Ser. 5, fasc. 3).

Leonhard Weber has, for a number of years, carried on a series of measurements of daylight at Kiel.[1] He has determined the noon illumination of a horizontal surface on the basis of a unit of light (*i.e.*, the quantity of light from a normal candle at the distance of 1 m.). This he has done for the green and the red rays, because white daylight cannot, from a physical standpoint, be strictly compared with candle-light. The mean of three years' observations showed that, taking the average of all the days, including clear, cloudy, and rainy, the daylight at noon was at a minimum in December. The values found were, for the red (in thousands of normal candles at the distance of 1 m.) 2·5, and for green, 9·0. The maximum noon daylight came in May and July, with 27·8 and 26·3 for the red, and 98·5 and 100·4 for the green, respectively. The mean cloudiness was 7·8, 6·0 and 7·1; the daily number of hours of bright sunshine 6·0, 7·9 and 7·1. The annual means are 16·3 for red and 59·4 for green. The total local illumination resulting from direct sunlight alone, on clear days, is reckoned by Weber at 3·4 in December for red and 5·4 for green; in May, 27·5 red, and 64·0 green; in July, 28·4 red, and 66·6 green. In the red, the total daylight is nearly equal to the computed values of direct sunlight alone. Cloudiness therefore does not change the value. In the green, the total daylight exceeds the direct sunlight. On days with white, bright clouds the amount of diffuse light can be very much increased. On June 5, 1892, the observed luminous intensity in green (284,000 normal candles at 1 m. distance) was four times as great as the daylight resulting from the direct sun's rays alone. On this day, three-quarters of the light were due to the diffuse light and one-quarter to direct sunlight. The cloudiness was estimated at 0·7, but the sun itself was not obscured.[2]

The annual march of the total radiant heat received from sun and sky on a body freely exposed in the air at Paris is shown by the following figures. The possible amounts of heat which can be received at the latitude of Paris are also entered in the table. It appears from these data that the heat radiation in July is more than 15 times as large as in December, and that, for the mean of the year, Paris has but little more than one-half of the amount of radiation to which it is entitled by reason of its latitude. In winter, this proportion is but one-third. It is interesting to compare with the foregoing data the annual march of the chemical intensity of radiation, and therefore there have been included in the table the monthly means of this

[1] *Schriften des naturw. Vereins für Schleswig-Holstein*, Vol. X.; also L. Weber: "Intensitätsmessungen des diffusen Tageslichtes," *M.Z.*, II., 1885, 163-172, 219-224, 451-455.

[2] See also H. W. Vogel: "Messung der Helligkeit des Tageslichtes," *Annalen der Physik und Chemie*, LXI., 408 [*M.Z.*, XIV., 1897, 300-301]. Chr. Jensen: "Beiträge zur Photometrie des Himmels," *Schrift. naturwiss. Ver. für Schleswig-Holstein*, XI., No. 2, 1899, 282-346 (Extract printed in *M.Z.*, XVI., 1899, 447-456, 488-499). C. Wiener: "Die Helligkeit des klaren Himmels und die Beleuchtung durch Sonne, Himmel und Rückstrahlung," *Abh. Kais. Leop.-Carol. Akad. d. Naturf.*, *Nova Acta*, LXXIII., No. 1, 1900 (*M.Z.*, XVIII., 1901, 43-47).

element as determined by Marchand, on the basis of four years measurements, for a station in the neighbourhood of Paris. At Fécamp, as at Kew, the chemical intensity of radiation is about ten times as great at the summer solstice as it is at the winter solstice. It is to be noted that all the data given in the table are to be considered as relative values only. The total chemical intensity of radiation from the sun and sky upon a horizontal surface at Fécamp, at the summer solstice, was equivalent to the production of 35 cu. cm. of CO_2. Of this amount, 23 cu. cm. were due to the sun alone and 12 cu. cm. to the diffuse skylight. A thin veil of cloud may increase fourfold the effect of the skylight.[1] Thanks to the measurements prompted by Bunsen and Roscoe, we now have considerable material for the determination of the chemical intensity of radiation in different climates, although the data at hand are still far from complete. With the sun's altitude nearly the same (53°) in all cases, the measurements gave the following relative chemical intensities of the total radiation from sun and sky: Manchester (53·5° N.), 183; Heidelberg (49·4° N.), 437; Para (1·5° S.), 724.

The mean chemical intensities of radiation at Kew were as follows: November—January, 11·0; February—April, 45·9; May—July, 91·5; August—October, 74·0; mean for the year, 55·3. Daily measurements of the total chemical intensity of daylight at St. Petersburg, at 1 P.M., gave a similar relation between winter and summer; the mean of the months November—February, was 0·03; and that of the months May—June, 0·26.[2]

Simultaneous measurements on three days in April, 1866, at Kew, near London, and at Para, gave a chemical intensity of radiation nearly twenty times greater at the latter station than at the former. If a

[1] For information concerning the photometer used by Marchand, see *M.Z.*, XIV., 1879, 258-259.

[2] The Arago-Davy actinometer used in Paris allows only the measurement of the solar rays of shorter wave-length than 3μ, the invisible portion of the spectrum through a large part of the infra-red being almost entirely excluded by the glass enclosure. This fact also appears in the value of the coefficient of absorption based upon the results obtained with this actinometer. This value was found to be 0·875, which is larger than Bouguer found it to be in the case of light. See also J. Wiesner: "Untersuchungen über das photochemische Klima von Wien, Cairo, und Buitenzorg (Java)," *D.W.A.*, LXIV., 1897, 73-166 [*Rev. M.Z.*, XIV., 1897 (24)-(27)]; "Beiträge zur Kenntniss des photochemischen Klimas im arktischen Gebiete," *ibid.*, LXVII., 1899, 643-676 [*Rev. M.Z.*, XVI., 1899, 525-526], and "Untersuchungen über den Lichtgenuss der Pflanzen im arktischen Gebiete," *S.W.A.*, CIX., 1900, 371-439.

TOTAL HEAT RECEIVED FROM SUN AND SKY AT MONTSOURIS.

Monthly Means Based on Eight Years' Observations.

Period	Dec.	Jan.	Feb.	Mar.	April.	May.	June.	July.	Aug.	Sept.	Oct.	Nov.	Year.
1872-1879	9·4	11·7	14·4	25·9	35·3	43·1	46·6	47·5	39·4	31·2	20·8	12·4	28·2
Maximum Possible Amount of Heat.													
	31·3	34·4	41·1	51·1	66·7	74·3	76·7	75·6	69·9	57·2	43·3	36·1	54·8
Mean Daily Chemical Activity at Fécamp, according to Marchand.													
1869-1872	1·8	1·8	3·9	6·4	14·1	19·5	21·0	21·4	18·9	13·7	6·9	2·9	11·0

comparison be made between August, at Kew, and April, at Para, the chemical activity of radiation at the latter place is still 3·3 times more intense than at the former.

Direct measurements of the chemical intensity of daylight and of sunlight, by Bunsen and Roscoe, make it possible to determine the dependence of these elements upon the sun's altitude, and thus to compute these intensities for different latitudes, with a close approximation to the actual conditions. The following figures give the computed relative chemical intensities of the radiation received directly from the sun and from the sky, upon a horizontal surface, during a whole day at the spring equinox. These figures emphasise very clearly the great advantage which equatorial climates enjoy in the matter of the intensity of direct solar radiation. These data may also be taken as giving an approximate measure of the mean annual chemical intensity of radiation, as far as the middle latitudes, because the greater intensity in summer very nearly compensates for the lessened intensity in winter; and the mean for the year thus comes back to the measure of the radiation at the equinox.

CHEMICAL INTENSITIES OF SUNLIGHT AND OF SKYLIGHT.

Place.	Latitude.	Chemical Intensity.		Total.
		Of Sunlight.	Of Skylight.	
Pole, - - - - -	90°	0	20	20
Melville Island, - -	75	12	106	118
Reikiavik, - - -	64	60	150	210
St. Petersburg, - -	60	89	164	253
Manchester, - - -	53	145	182	327
Heidelberg, - - -	49	182	191	373
Naples, - - - -	41	266	206	472
Cairo, - - - -	30	364	217	581
Bombay, - - -	19	438	223	661
San José (Costa Rica), -	10	475	226	701
Quito, - - - -	0	489	227	716

It is also apparent from the foregoing table that the diffuse skylight exceeds direct solar radiation as far as the latitude of Heidelberg, and very considerably exceeds it in higher latitudes. At the equator, however, the intensity of the former is hardly over one-half of the

direct solar radiation. This emphasises very clearly the important part played by the atmosphere in regulating the distribution of the total radiation, by lessening the great differences in direct radiation at different latitudes. Heidelberg, for example, would receive from the sun alone but about one-third, and St. Petersburg less than one-fifth, of the directly received and chemically active radiation at the equator. As the result of the dispersive power of the atmosphere, however, this ratio is increased to more than one-half, and more than one-third, respectively.[1]

Thermal, optical, and chemical climatic zones.—Bunsen and Roscoe call attention to the important fact that the photochemical, as well as the optical climatic belts differ essentially from the thermal climatic regions in that the actual distribution of the former must be governed by a much simpler law than that of the latter. The temperature of the air, which is primarily due to insolation, is irregularly distributed over the earth's surface by aerial and oceanic currents, because the heat produced at any place may be carried elsewhere by convectional currents, as also in the form of the latent heat of water vapour. This, however, is not the case with the chemical activity of the sun's rays, since this effect of solar radiation must remain exactly where the direct sun's rays fall. The optical, chemical, and, we may add, the heating effects of direct radiation, as contrasted with air temperature, are limited in their distribution by parallels of latitude. There can then be no doubt that direct solar radiation is one of the most important climatic factors. For this reason, the study of the distribution of the total intensity of solar radiation is of much more consequence in climatology than is generally supposed. We must therefore agree that the climatic zones which were adopted by the ancient geographers are of value, in so far as they represent the principle of classifying climatic districts according to latitude.[2]

[1] Ed. Stelling: "Photochemische Beobachtungen der Intensität des gesammten Tageslichtes in St. Petersburg," *Repertorium für Meteorologie*, VI., No. 6, and *Z. f. M.*, XIV., 1879, 41-48. See also J.M. Pernter: "Resultate der bisherigen photochemischen Messungen des Sonnenlichtes," *Z. f. M.*, XIV., 1879, 401-426.

[2] The early studies of solar radiation, up to the year 1875, are admirably compiled and discussed by R. Radau: *La Lumière et les Climats. Les Radiations chimiques du Soleil. Actinométrie.* These were all published in 1877 by Gauthier Villars, in Paris. See also P. Houdaille: *Météorologie agricole. Le Soleil et l'Agriculteur.* Masson, Paris, 1893. The following may also be found of interest:—M. Andresen: "Influence de la Pression Barométrique sur l'Action Chimique de la Lumière directe du Soleil," *Ann. Obs. Met. Mont Blanc*, IV., 1900, 1-17.

Selective absorption by the atmosphere, and its relation to temperatures on the earth's surface.[1]—If we would understand, and appreciate, the important part played by the atmosphere in the thermal economy of the earth, we must, in conclusion, consider at somewhat greater length the relation of the earth's atmosphere to solar radiation, in the light of the most recent studies. As the result of a long series of very careful determinations of the absorption of the different solar rays, Langley has found that the relative absorption decreases from the blue end of the spectrum towards the red end. Hence the so-called "chemical" rays, and also the luminous rays, suffer a greater relative loss in their passage through the earth's atmosphere than do the red and the shorter infra-red rays. Langley has thus shown that the view which was formerly very commonly held, that the atmosphere is, like glass, almost impassable by the totality of the infra-red rays, is an erroneous one.[2] The extreme infra-red rays beyond 5μ are, however, almost totally absorbed as far as 8μ, and other extensive gaps exist.

With increasing altitude of the sun, and hence a decreasing path through the atmosphere, the sunlight becomes much richer in blue, short-waved, more refrangible rays. The maximum intensity of radiation in the spectrum is between red and orange (wave-length 0·7 microns) when the sun is low; in the middle of the yellow ($\lambda = 0{\cdot}6$) when the sun is high; and probably between green and blue ($\lambda = 0{\cdot}5$) at the upper limit of the atmosphere.

The theoretical studies of Lord Rayleigh have made clear the nature of this effect which the atmosphere has upon the radiant energy of the sun. Rayleigh has shown that solar radiation, made up, as it is, of many different waves, in passing through a turbid medium like our atmosphere, undergoes diffuse reflection and scattering, and is thereby weakened in such a way that the short-wave rays are most, and the long-wave rays are least affected by the process.[3] This agrees with the fact discovered by Langley, above referred to, that there is a decreasing coefficient of

[1] See S. P. Langley: "Researches on Solar Heat and its Absorption by the Earth's Atmosphere," *Professional Papers of the Signal Service*, No. XV., 4to, Washington, D.C., 1884, pp. 242; C. G. Abbot: "The Relation of the Sun-spot Cycle to Meteorology," *Monthly Weather Review*, XXX., 1902, 178-181.

[2] According to Langley, the coefficients of transmission, when the sun is high, are 0·35 for rays at the violet end; 0·88 for those in the middle of the yellow, and 0·95 for those at the red end of the spectrum.

[3] The amount of diffuse reflection is inversely proportional to the fourth power of the wave-length.

transmission from the red towards the blue end of the spectrum. Abney was afterwards able to show by direct observation that the law deduced by Rayleigh, as to the diffuse reflection from fine particles suspended in a medium, suffices to explain the observed weakening of the transmitted sunlight. Very has further recently published an important paper on atmospheric radiation[1] in which radiation by the atmosphere, and by other gases, especially carbon dioxide and water vapour, as well as the absorption of radiation by these substances, are discussed.

The incoming solar radiation has suffered depletion by the absorbent action of various atmospheric constituents, as well as by a selective diffuse scattering of the shorter waves by the air molecules, and by the fine suspended dust and water particles of the lower air. The absorption by carbon dioxide diminishes the solar radiation by about $1\frac{1}{2}$, or at most 2 per cent. This loss is sensibly constant, at least for ordinary elevations above sea level, since the amount of the gas in the atmosphere, although small, is always more than sufficient to produce almost total repression of those special radiations which are the only ones capable of being cut off by this substance.

The absorption exercised by aqueous vapour varies both with the absolute and the relative humidity, and is very local in its action through a large part of the infra-red spectrum, producing a series of bands and groups of fine lines, which culminate between the wave-lengths of 5 and 8 microns. Abney and Festing[2] find that, over the fine telluric lines produced in the solar spectrum by water vapour, there are superimposed other diffuse bands of great intensity, when the air is nearly saturated with moisture. Very[3] attributes the diffuse bands to complex aqueous molecules which form in the air as saturation approaches, and which greatly increase the absorptive power of the nearly saturated vapour. If f is the tension of aqueous vapour in mm., and h, the fraction expressing relative humidity, W, the absorption of solar radiation by water vapour, is approximately given by the empirical formula:[4]

$$\log W = -\frac{0{\cdot}999\ldots}{1+\log(1+fh)^{1{\cdot}65}}.$$

The band absorption (B) and the depletion by selective scattering from fine particles must be treated separately. Very finds it convenient to divide the latter into scattering by air molecules (transmission factor R), and scattering by dust (transmission factor D), so that if d and d' are the paths through the

[1] F. W. Very: "Atmospheric Radiation," *U.S. Dept. of Agriculture, Weather Bureau, Bulletin G*, 4to, Washington, D.C., 1900, pp. 134.

[2] Abney and Festing: "Atmospheric Absorption in the Infra-red of the Solar Spectrum," *Proc. R. S. London*, XXXV., 1883, 80-84.

[3] F. W. Very: "Atmospheric Radiation," pp. 100-103.

[4] F. W. Very: "The Solar Constant," *U.S. Weather Bureau*, No. 254, 1901, pp. 29.

aerial and the dust envelopes respectively, A, the solar constant, and I the observed solar radiation :

$$I = (A - B) \times R^{d} \times D^{d'}.$$

The quantity I given by the formula on p. 104 is not the true solar constant, since a large part of solar radiation disappears by band absorption in the upper air, and is lost to observation. The immediate effect of this high-level absorption is to raise the level of the altitudinal isotherms, which, in turn, reacts upon surface temperature, and prevents sudden excessive chilling of the surface by descending air currents, because the protective layer of warm, moist, and therefore highly absorbent air is thicker.

The colour of the sky.—The light and heat, which are thus diffusely scattered, are by no means lost to the earth's surface, for they illuminate the sky ; they give us the diffuse daylight, and make the whole firmament a source of light and of warmth. Since the blue rays undergo the greatest amount of scattering, therefore the blue of the sky is explained, as is also the larger proportion of blue, chemically active, rays in the diffused light.[1]

Atmospheric absorption of terrestrial radiation.—It was customary, up to the time of recent discoveries, to compare the effect of the atmosphere upon temperatures on the earth's surface to that of the panes of glass in a hot-house. The glass is virtually transparent to the luminous rays from the sun ; but it is almost wholly impassable by the invisible rays of long wave-length emitted from bodies which have been warmed by the sun within the hot-house. Thus the heat is as it were imprisoned within the glass. The rays of short wave-length which form the principal part of solar radiation were admitted freely, but being converted within the hot-house into rays of long wave-length, they can no longer pass outward through the glass. This explanation was at first apparently overthrown by Langley,[2] whose earlier investigations with a glass prism showed that the red and the infra-red rays are in general less absorbed by the atmosphere than the luminous rays of short wave-length. Later observations, made with rock-salt prisms, however, have disclosed extensive spectral regions of still greater wave-length, in which great lacunae, caused by the band absorptions due to water vapour and carbon dioxide, restore the

[1] See a recent paper by N. E. Dorsey : "On the Color and Polarization of Blue Skylight," *Monthly Weather Review*, XXVIII., 1900, 382-389. A bibliography is appended to this article.

[2] *Annals Astrophys. Obsy. Smiths. Inst.*, I., 1900, contains Langley's recent work on the lower infra-red spectrum. See *Monthly Weather Rev.*, XXX., 1902, pp. 258-260.

analogy. It is therefore admissible to compare the atmospheric envelope of the earth with the glass windows of a hot-house.

Beside the effect of the atmosphere upon the solar radiation passing through it, which has been discussed above under the term *diffuse reflection*, there is still another effect. The atmosphere wholly *absorbs* certain groups of rays, with the result that they appear in the spectrum as dark bands, comparable to the Fraunhofer lines of the sun's atmosphere. There are certain broad, cold bands, especially in the infra-red portion of the spectrum, which prove that these missing waves have been entirely absorbed by the earth's atmosphere, and that the atmosphere must therefore have become warmed at the expense of these rays. As these broad, cold bands are found chiefly in the infra-red portion of the spectrum, the atmosphere does present some analogy with glass. From this point of view, therefore, the effect of the atmosphere upon temperatures at the earth's surface, may perfectly well be compared with the effect of the glass windows of the hot-house. Special rays of long wave-length, emanating from the earth's surface which has been warmed by the luminous solar rays, are very largely absorbed and held back by the atmosphere. Thus heat is stored in the atmosphere, as it is behind the glass windows of a hot-house.

The absorption of the invisible rays by the atmosphere has been shown by Angström and Paschen to be due chiefly to the water vapour and carbon dioxide in the air. Oxygen and nitrogen show scarcely any such effect. Notwithstanding the small proportion of carbon dioxide in the atmosphere, its effect upon the heating of the air cannot be disregarded.[1]

[1] Lecher has stated that a stratum of CO_2, one meter in thickness, absorbs 13 per cent. of the radiation from the sun. In view of this absorption, Knut Angström was formerly inclined to assume a higher value for the solar constant than that given by Langley, placing it as high as 4 calories (Wiedemann's *Annalen*, XXXIX., 1890, 267, 294). Very has pointed out that Angström's method implies that 60 per cent. of the original solar radiation is included within the limits of the CO_2 bands, but actually only a small fraction is contained within these limits, whence the value of 4 calories for the solar constant is inadmissible on these grounds (*Atmospheric Radiation*, 105). Angström himself has since proved that Lecher's observation was erroneous, and has withdrawn his higher estimate of the solar constant. See also :—S. Arrhenius : "Ueber den Einfluss des atmosphärischen Kohlensäuregehalts auf die Temperatur der Erdoberfläche," *Bih. till K. Sv. Vet-Akad. Handl. B.*, Vol. XXII., 1, Stockholm, 1896 ; Nils Ekholm : "On the Variations of the Climates of the Geological and Historical Past and their Causes," *Quart. Journ. Roy. Met. Soc.*, 1901, 1-62 ; Knut Angström : "Intensität der Sonnenstrahlung in verschiedenen Höhen nach Untersuchungen auf Tenerifa, 1895 und 1896," *Naturwiss. Rundschau*,

Effect of water vapour upon the absorption of radiation by the atmosphere.—The chief part in the absorption of radiation by the atmosphere is undoubtedly played by the water vapour in the atmosphere. A much greater amount of heat can be stored at the earth's surface in an atmosphere which contains much water vapour (and carbon dioxide) than in a very dry atmosphere. As the amount of water vapour in the air decreases with altitude much more rapidly than the pressure, the influence of the atmosphere in absorbing and storing terrestrial radiation decreases rapidly with increase of altitude. The absorption of solar radiation, however, both by water vapour and by carbon dioxide, takes place chiefly in the upper air, because small amounts of these absorbents, when acting upon an undepleted solar radiation, are quite effective; but after the more absorbable rays have been taken out, the remaining ones pass comparatively freely, even through layers of absorbing vapour of much greater density. The presence of water vapour is seen, by the active absorption it produces, to be effective almost wholly in the infra-red portion of the spectrum, and hardly at all in the optical portion. The losses which solar radiation suffers, as regards its lighting and heating qualities, therefore do not vary at equal rates, as Langley has already shown. Thus it is seen that the heating effects of solar radiation decrease much more rapidly with increasing humidity of the atmosphere than the lighting effects.

Searle has made some valuable suggestions in connection with the atmospheric economy of solar radiation.[1]

The atmosphere, in the first place, has the property of easily acquiring heat from the warmed surface of the earth, because the particles of air which are warmed below ascend and make way for colder particles. Thus heat is carried up into the upper strata by means of convectional currents. On the other hand, this heat is not so easily lost, even when the earth's surface cools; for then the cooling acts only upon the lowest strata of air by conduction. Hence the return of heat from the atmosphere to the earth is a comparatively slow and imperfect process. A check upon the process of accumulation of the stock of energy in the atmosphere is found in the results of the movement of the air as wind. When the winds have attained a certain degree of violence, they disturb the portions of the air which would otherwise remain stagnant over the colder parts of the ground, and the

XV., 1900, 649; "Ueber die Bedeutung des Wasserdampfes und der Kohlensäure bei der Absorption der Erdatmosphäre," *Annalen d. Physik*, 4te Folge, III., 1900, 720-732 (*M.Z.*, XVIII., 1901, 470-471); "Ueber die Abhängigkeit der Absorption der Gase besonders der Kohlensäure," *ibid.*, V., 1901, 163-173. See also: "Knut Angström on 'Atmospheric Absorption,'" *Monthly Weather Review*, XXIX., June, 1901, 268.

[1] Arthur Searle: *Proc. Amer. Acad. Arts and Sci.*, XVI. (new series), 1888-89, 26-29.

process of heating the atmosphere from beneath gradually ceases to retain sufficient advantage over that of cooling it from beneath to permit a further accumulation of energy. The atmosphere thus acts as a check upon extreme variations of heat and cold. When little heat is received, it will be better economised than when the supply of heat is excessive, although particular regions may have, in the former case, a very severe climate. There is, furthermore, another consideration. The warmed, ascending air cools by expansion, *i.e.*, dynamically, but its potential temperature remains the same. When this air again descends, as, for example, in the atmospheric circulation which takes it to higher latitudes, or as a foehn, it is again warmed, and its former warmth is again available at the earth's surface. The fact that this air, after its ascent, is at a much lower temperature, without at the same time sacrificing its potential temperature, protects it from any considerable loss of heat by radiation. Thus the gaseous, elastic envelope of the earth favours the collection of a certain store of heat in itself.[1]

[1] See also G. A. Hirn: "Introduction à l'Étude météorologique et climatérique de l'Alsace;" *Extr. du Bull. de la Soc. d'Hist. nat. de Colmar*, Colmar, 1870, 64.

APPENDIX TO CHAPTER VI.

DETERMINATION OF THE INSOLATION FACTORS.

If we express the declination of the sun by δ, the latitude by ϕ, the length of the semi-diurnal arc by t, the apparent semi-diameter of the sun by d, and use C to denote a constant value, which must be determined at the start, we obtain the total daily insolation on any given day in the year and at any given latitude, by means of the formula,[1]

$$\text{I.} \qquad W = Cd^2(\sin\delta \sin\phi t + \cos\delta \cos\phi \sin t).$$

The length of half the diurnal arc is computed by the well-known formula,

$$\cos t = -\operatorname{tang}\delta \operatorname{tang}\phi.$$

By substituting this equation in equation I., we have the latter in a more convenient form for purposes of computation, namely,

$$\text{II.} \qquad W = Cd^2 \sin\delta \sin\phi(t - \tan t).$$

The expression t is the length of the diurnal arc for a radius of 1. For the spring equinox and for the equator $\left(\delta = 0,\ \phi = 0,\ t = \frac{\pi}{2}\right)$, equation I. becomes

$$W' = Cd^2,$$

by means of which the constant C can be computed. Taking the value of $W' = 1000$ (Meech = 1), with $d = 965''$ (on March 20), $C = 0{\cdot}00107$. We may, however, assuming a given solar constant (A), also compute this quantity of heat in calories.

When the sun is on the equator, the day is 12 hours long, and the quantity of heat upon a surface always at right angles to the sun's rays is $12h \times 60m \times A = 720A$. The insolation upon the horizontal surface of the earth is naturally less, in the same ratio as that between the diameter of the semicircle (since the diurnal arc is perpendicular to the horizon), and

[1] See Meech: "On the Relative Intensity of the Heat and Light of the Sun," *Smithsonian Contributions to Knowledge*, IX., 1857, 7-58; or Wiener: "Ueber die Stärke der Bestrahlung der Erde durch die Sonne in verschiedenen Breiten und Jahreszeiten," *Z. f. M.*, XIV., 1879, 113-130.

the semicircle itself, *i.e.*, $2r : r\pi = 2 : \pi$. The quantity of heat is therefore $(2 \div \pi) \times 720A = 458 \cdot 4A$. When the solar constant (A) for the mean distance of earth from sun is substituted in this equation, and is, for example, following Langley, taken as 3, the quantity of heat at the equator on March 20 is $1375 \cdot 2 \times (965'' : 961'')^2 = 1386 \cdot 7$ calories. Above, we have taken this amount as 1000, which corresponds to a solar constant of 2·16 calories. Our relative values should therefore be increased by about 39 per cent. in order to give the amount in calories; while the values given by Angot should be increased by 37·5 per cent. Angot expressed the quantity of heat when the earth is at an average distance from the sun, *i.e.*, when $d = 961''$, by 1000.

Using units of heat obtained by the solar constant of Langley, equation II. becomes

III. $$W = 1386 \cdot 7 (d : 965'')^2 \sin\delta \sin\phi (t - \tan t).$$

If the sun were always on the equator, equation I. would give a simple law for the decrease of temperature with increase of latitude. For, if we have $\delta = 0$, t is everywhere $= \frac{\pi}{2}$, and the equation becomes

$$W = Cd^2 \cos\phi.$$

The heat would therefore decrease with the cosine of the latitude. As a matter of fact, this law is approximately true for low latitudes.

In the case of the quantity of heat at the pole during the time when the sun is above the horizon and does not set, we obtain from equation I., when $\phi = 90°$ and $t = \pi$, because half the diurnal arc then becomes 180°,

$$W = Cd^2 \pi \sin\delta.$$

As, in the case of the equator, when $\phi = 0$, and $t = \frac{\pi}{2}$,

$$W = Cd^2 \cos\delta,$$

the quantity of heat received simultaneously at the pole is to heat received at the equator, as $\pi \sin\delta : \cos\delta$, or W at the pole $=$ W at the equator $\times \pi \tan\delta$. The quantities of heat at pole and equator are equal when $1 = \pi \tan\delta$, or when $\delta = 17° 14'$, *i.e.*, on May 10 and August 3. During 86 days, therefore, the insolation at the pole is stronger than at the equator. When $\delta = 23° 27'$, the maximum insolation at the pole $=$ equator $\times$ 1·364. The insolation at the limit of the atmosphere at the pole is then 36 per cent. in excess of that at the equator at the same time.

For a special place, such as Vienna ($\phi = 48° 12'$), for example, equation III. gives, for the solstices,

1. December 21. $\delta = -23° 27'$, $d = 978''$.

$$t = 4h\ 5m = 61° 15',$$

or, in circular measure, 1·0690.

$$\tan t = 1 \cdot 8228\ ;\ \text{hence } t - \tan\delta = -0 \cdot 7538,$$

$$\text{and } W = 319.$$

2. June 21. $\delta = +23° 27'$, $d=946''$, $t=118° 45'$.

$$t - \tan\delta = 3{\cdot}8954, \text{ whence } W=1541.$$

For a station in the southern hemisphere, in the latitude of Vienna, these values would be 298 on June 21, and 1647 on December 21. These results are obtained directly from the values above, by multiplying by $(946 : 978)^2$, and *vice versa*.

If the sun were always at a mean distance from the earth, and over the equator, the quantity of heat which the equator would receive in the course of a year would be

$$365{\cdot}24 \times 458{\cdot}4 \times A = 167{,}416A.$$

As the sun, however, moves away from the equator by the amount of the obliquity of the ecliptic, the annual quantity of heat which the equator receives is less in the ratio by which the circumference of an ellipse with a semi-major axis a, and an eccentricity equal to the sine of the obliquity of the ecliptic, is smaller than the circumference of a circle with a radius a. The factor for 23° 28′ is therefore 0·95918; whence the annual quantity of heat at the equator is: $160{,}582 \times A$; or, if $A=3$, the annual radiation at the limit of the atmosphere should give 481,750 calories.

As $365{\cdot}24 \times 0{\cdot}9592 = 350{\cdot}4$, Angot places the quantity of heat at the equator as equal to 350·4 mean "heat days." This number must be multiplied by $458{\cdot}4A$, in order to give the quantities of heat in calories, while the values given in our table, on p. 100, need to be multiplied by $439{\cdot}7A$ only.

The following table gives a few of the insolation factors computed by Angot for different latitudes and for various coefficients of transmission. This table makes no allowance for cloudiness, nor for the radiation extinguished by band absorption. It therefore gives values which are always larger than those observed. In general, as has been noted in regard to the observations at Paris, the values computed by the aid of the insolation factors are about three times too large. In this table "summer" means the period of the northern declination of the sun, and "winter" the period of southern declination. The maxima and minima are the quantities of heat from radiation on the extreme days. The figures given for summer, winter, and the year are, however, heat days in the above sense, and hold for the southern hemisphere as well.

INSOLATION FACTORS AT DIFFERENT LATITUDES OF THE NORTHERN HEMISPHERE, AND WITH DIFFERENT CO-EFFICIENTS OF ABSORPTION BY THE ATMOSPHERE

(AFTER ANGOT).

Coefficient of Transmission.	1	0·8	0·7	0·6	1	0·8	0·7	0·6
		Equator.				40° N.		
Maximum,	1010	737	617	506	1115	796	660	536
Minimum,	888	629	517	416	360	184	124	84
Summer,	175·1	126·0	104·6	85·1	185·2	129·8	106·6	85·6
Winter,	175·1	126·0	104·6	85·1	91·6	54·3	40·6	29·6
Year,	350·3	251·9	209·2	170·2	276·8	184·1	147·2	115·2
		10° N.				50° N.		
Maximum,	1010	734	614	504	1112	771	633	507
Minimum,	825	566	459	368	198	73	41	22
Summer,	185·4	134·3	112·2	91·8	175·5	118·5	95·6	75·2
Winter,	160·1	113·0	92·9	74·7	64·1	32·7	22·9	15·4
Year,	345·5	247·3	205·1	166·5	239·6	151·2	118·5	90·6
		20° N.				60° N.		
Maximum,	1052	763	639	523	1101	730	585	456
Minimum,	681	442	349	267	56	5	1	0
Summer,	190·4	137·7	115·1	94·1	162·6	103·6	81·0	61·8
Winter,	140·8	95·9	77·5	61·0	36·6	14·6	9·2	5·6
Year,	331·2	233·6	192·6	155·1	199·2	118·2	90·2	67·4
		30° N.				70° N.		
Maximum,	1096	794	661	541	1137	680	525	393
Minimum,	524	312	236	171	0	0	0	0
Summer,	190·3	136·4	113·1	92·0	151·4	86·2	64·4	46·5
Winter,	117·6	75·9	59·8	45·6	14·8	4·4	2·4	1·2
Year,	307·9	212·3	172·9	137·6	166·2	90·6	66·8	47·7
		80° N.				90° N.		
Maximum,	1192	685	497	347	1210	691	494	335
Year,	150·2	73·6	50·5	33·5	145·4	68·2	45·0	28·4

SECTION II.—THE CHIEF VARIETIES OF CLIMATE AS MODIFIED BY THE SURFACE FEATURES OF THE EARTH—PHYSICAL CLIMATE.

INTRODUCTION.

PHYSICAL CLIMATE.

Continental, marine and mountain climates.—Solar climate, as modified by the surface features of the earth, is usually called *physical*, or *natural climate*. These surface features react upon the atmosphere, and thus interfere with the uniform arrangement of the climatic zones, and with the simple demarkation by parallels of latitude which would exist in a purely solar climate. The chief causes of this interference with the regular solar climatic zones are (*a*) the irregular distribution of land and water upon the earth's surface, (*b*) the aerial and oceanic currents, which are thereby compelled to follow certain definite paths, and (*c*) the difference in altitude of the land above sea-level. These factors determine the two chief classes of climates found on the earth, viz., I., Continental and marine climates; and II., Mountain climates.

The fact that a parallel of latitude runs partly over land and partly over water brings about differences in climate between east and west which would not exist in the case of solar climate alone. In addition to the unequal warming and cooling of land and water, certain prevailing currents are caused by the presence of the land, which themselves give rise to differences in climate along the same parallel of latitude. If the earth's surface were uniform, the atmospheric and oceanic currents produced by the interchange of heat between equatorial and polar regions would not be limited to certain meridians. As a matter of fact, however, the presence of land and water surfaces, and especially their extension along the meridians, determine that

these currents shall prevailingly follow definite paths. Hence it comes about that the transfer of heat from lower to higher latitudes proceeds chiefly along some meridians, while the transfer of cold from higher to lower latitudes follows certain other meridians.

Let us now examine more closely the different influences of land and of water upon climate, as well as the influence of the land upon the direction of the prevailing currents in the atmosphere and in the oceans.

A. CONTINENTAL AND MARINE CLIMATES.

CHAPTER VII.

INFLUENCE OF LAND AND WATER UPON THE DISTRIBUTION OF TEMPERATURE.

Specific heat of land and of water.—Land and water differ in their behaviour regarding the gain and loss of heat by radiation, and these two processes are the chief factors upon which the air temperature at any place depends. The specific heat of water is higher than that of any other substance upon the earth's surface, and has been taken as the unit of specific heat. The specific heat of a land surface may be taken as about equal to 0·2, when equal weights of water and of dry soil are compared. When, however, equal volumes are compared, and this is the case with which we have here to deal, the specific heat of the land is about 0·6 of that of water. This is equivalent to saying that when equal quantities of heat are received by equal areas of land and of water, the resulting increase of temperature is almost twice as great on the land as on the water, even when, in the case of the water, no considerable share of the heat received is expended in the process of evaporation.

Effect of evaporation upon the heating of water.—The increased evaporation which results from the heating of water lessens the increase in the temperature of the water to a marked degree. The greater portion of the heat developed from the radiation received by water is probably expended in the production of water vapour, and a small fraction only remains to be employed in raising the temperature of the water.

Fitzgerald has made some observations on the daily warming of Lough Derg in clear, hot summer weather. The surface temperature rose during the morning

at the rate of almost 0·6° an hour. "From a calculation of the amount of heat that enters the water, it seems that only about 1/50, or less, was used in heating it, the rest being probably spent in evaporation." From the rate of decrease of temperature in the superficial layers, Fitzgerald calculated that the coefficient of absorption of heat per yard (0·9 m.) was ·71.[1]

The annual evaporation at the equator may, on the basis of trustworthy measurements, be assumed to be 2·3 m.[2] Assuming a mean water temperature of 27°, this requires an amount of heat equivalent to about 135,000 calories. As the maximum amount which the equator can receive, assuming the absorption to be that in a perfectly clear atmosphere, is barely 240,000 calories, it appears that nearly 0·6 of this amount is expended in evaporation. On account of the cloudiness of the sky, the amount of available energy remaining to heat the water must be even considerably less.

Diurnal and annual march of temperature in water and underground.—Water is thus seen to warm more slowly, and to a less degree, than land, because of its high specific heat, and especially because of the evaporation which accompanies its warming. Moreover, there is still another factor which is concerned in this process. The daily changes in temperature penetrate into the ground to a depth of about 1 meter only, while they reach depths of 10 to 20 meters in the water. Solar radiation penetrates the water,[3] and warms even the deeper strata directly. In the case of the land, however, the heat from the surface can reach the lower strata only by conduction. Water is therefore heated to a greater depth than land, but is correspondingly less warmed on the surface. The annual variation in temperature is no longer perceptible at a depth of 20 meters in the ground, while it reaches a depth

[1] G. F. Fitzgerald : "On the Temperature at Various Depths in Lough Derg after Sunny Weather," *Sci. Proc. Roy. Dub. Soc.*, V., 1886-87, 169-170 [*M.Z.*, XVIII., 1888 (22)].

[2] It is customary to overestimate the amount of evaporation in warm climates, for the reason that the measurements are, in most cases, not made under natural conditions. Some observations made at Adelaide, South Australia, which are trustworthy in this respect, show an evaporation of only 140 cm. a year, while the evaporation at Alice Springs, in the hot, dry interior of Australia, is given as only 258 cm., the rainfall in this district being but 29 cm. In July, 1876, Dieulafait found the daily evaporation of ocean water, 15 km. from the coast, between 11 and 12 mm. during a calm. In another place, during July and August, the amount was observed to be 8 to 13 mm. Dieulafait estimates the mean annual evaporation over the open sea, on the Mediterranean coast of France, at 2·2 m. Pechinet gives the annual evaporation in the Camargue as 2·5 – 2·7 m. ("Evaporation de l'Eau de Mer dans le Sud de la France et en particulier dans le Delta du Rhône," *Comptes Rendus*, XCVI., 1883, 1787-1790).

[3] The red and the infra-red rays are the most absorbed ; the luminous rays, less absorbed ; and the blue rays are the most scattered.

of 100 to 200 meters in oceans and seas. The reason for this is found in the presence of ascending and descending currents in the water. The water which is cooled on the surface becomes heavier and sinks, while warmer water rises to take its place. In the case of oceans which contain a considerable amount of salt, the evaporation at the surface makes the surface water more saline and heavier. This water, with its higher temperature, therefore sinks to greater depths, and thus the lower strata can be warmed to the temperature of the surface. The movement of the water in waves, and the mixture of the different strata by the winds, also play a part. During the summer, a considerable quantity of heat is thus stored in the larger bodies of water, and this heat is slowly given out again during the winter. The store of heat in the dry ground is very much smaller in comparison, and is quickly given out again in the autumn, or at any time of falling temperature.

Surface temperatures of oceans and lakes.—The mean surface temperature of rivers, lakes and oceans is higher than that of the air over them. This is true to a marked degree of enclosed bodies of water in warm climates. On the basis of all available observations, Köppen did not find the mean annual temperature of the open oceans notably higher than the air temperature. In the case of the Atlantic Ocean, from latitude 10° S. to 20° N., the surface temperature was only 0·2° above that of the air. The water of warm ocean currents was in general, for the mean of the year, about 1° warmer than the air. The air over cold currents, however, was hardly any warmer than the water.[1]

Schott finds a much greater excess of temperature in the water, a result probably due to the fact that he took his own observations very carefully with the Assmann aspiration psychrometer.[2] According to Schott's observations, the open oceans in the tropics are 0·8° warmer than the air over them, while the open extra-tropical oceans are actually 1·6° warmer. In mediterranean seas, this difference amounts to 2°, or more. The Lake of Geneva has a surface temperature 2·4° above that of the air over it. This difference is +4·8° in winter;

[1] Cf. W. Köppen: "Ueber das Verhältniss der Temperatur des Wassers und der Luft an der Oberfläche des Ozeanes," *Annalen der Hydrographie*, XVIII., 1890, 445-454.

[2] The result of a comparison of Schott's observations with those made in the usual way, shows that air temperatures obtained by the ordinary method average about 0·6° too high at sea in the tropics. ("Wissenschaftliche Ergebnisse einer Forschungsreise zur See, 1891 und 1892." *Petermann's Mittheilungen*, Ergänzungsheft 109, 1893.)

$-0{\cdot}3°$ in spring; $+1{\cdot}3°$ in summer; and $+3{\cdot}9°$ in autumn. It is thus clear that considerable bodies of water which do not freeze in winter are an important source of warmth for places in their vicinity during the colder months. It results further from the conditions of warming of water as compared with land that, in spring, with rising temperature, the water is colder than the land, and therefore has a cooling influence upon its surroundings. In the autumn, when the temperature of the air falls, the water is warmer than the air, and therefore has a warming effect upon its surroundings.

Walter has compared the mean temperature of stations on the shore of Lake Constance, and of other stations farther from, but in the neighbourhood of the lake, the temperatures all being reduced to the same altitude above sea-level (400 m.). The results show that the mean annual temperature on the shore of the lake is 8·6°, while away from the lake it is only 8·2°, or 0·4° lower. The shore is 0·8° warmer in January, and there is no difference in March and April. From August to September, the lake shore is again warmer by 0·6° – 0·7°. The differences in temperature between the spring and the fall months are as follows:—

	Sept.—May.	Oct.—April.	Nov.—March.
Lake shore, - -	+1·1°	–0·1°	+0·9°
Inland, - - -	+0·6	–0·7	+0·4
Excess on the Lake,	+0·5	+0·6	+0·5

The autumn is found to be half a degree warmer on the lake shore.[1] The Lake of Geneva has a similar effect upon places in its vicinity. Vevey and Montreux are 1° warmer than Bex in fall and winter.[2] Petterson[3] has recently brought forward some striking evidence as to the effect of the temperature of the ocean surface upon the temperature of the air. He finds, among other things, that every square meter of the surface water of the North Sea gives off to the air above it 150,000 large calories of heat from August to

[1] *Eine Studie ueber die Temperatur- und Niederschlagsverhältnisse im Bodensee-becken*, Freiburg i. Br., 1892.

[2] Bührer: "Le Climat du Canton de Vaud," *Archiv Sci. phys. et nat.*, Geneva, 1896, I., 372-375.

[3] O. Petterson: "Beziehungen zwischen hydrographischen und meteorologischen Phänomenen," *M.Z.*, XIII., 1896, 285-321.

November, and 540,000 calories from November to February. In the Baltic Sea, the corresponding values are 130,000 and 355,000 calories.

The difference in the behaviour of land and water is well shown in a series of simultaneous observations of soil temperatures and of the temperature in a lake at Pawlowsk, near St. Petersburg, in June, 1882.[1]

	Surface.	0·01 m.	0·20 m.	Difference.
Air Temperature, Mean,	14·8°	—	—	—
Soil Temperature, Mean,	19·5	19·3°	15·8°	3·7°
Water Temperature, Mean,	19·8	19·7	19·3	0·5
Soil Temperature, 1 p.m.,	28·7	27·9	16·1	12·6
Water Temperature, 1 p.m.,	22·5	21·8	20·7	1·8

The marked excess of warmth in the water is noticeable at even the slight depth of 2 dm. The extreme temperatures at the surface were as follows: Air, 3·9° and 26·4°; soil, 5·4° and 44·8°; water, 13·5° and 29·8°. This gives a range of 22·5° in the air; 39·4° in the ground; and 16·3° in the water.

The diurnal and annual changes in the temperature of the ocean surface are small, and for this reason the oceans have an important influence in diminishing the diurnal and annual ranges of temperature on coasts and on islands. The mean diurnal variation of temperature over the open ocean within the tropics hardly reaches 1°, as was observed by Humboldt many years ago. The annual variation in temperature is also very small in low latitudes; and the times at which the maximum and the minimum occur are considerably retarded. Thus, in the northern hemisphere, the minimum occurs in February, or even in March, while the maximum occurs in August and September. The Atlantic Ocean, in latitude 35° N. and longitude 0° – 50° W., has a temperature of 16·7° in February, and of 24·0° in August, the annual range being 7·3°. The annual range in the temperature of the ocean water has been made the subject of a special investigation by Schott, who has also charted the result of his observations.[2] He has established the interesting fact that, with the exception of the marginal portions, all the oceans have the smallest annual range of temperature in the equatorial regions, and that this range increases to the north

[1] A. Woeikof: "Klimatologische Zeit- und Streitfragen." *M.Z.*, V., 1888, 205-211.

[2] G. Schott: "Die jährliche Temperaturschwankung des Ozeanwassers," *Pet. Mitt.*, XLI., 1895, 153-159.

and south as far as latitude 30° to 40°, from which point it again decreases. The accompanying table contains some general mean temperatures for the oceans, and also presents a comparison with the annual ranges of temperature on the continents (after Supan).

ANNUAL RANGES OF TEMPERATURE OF OCEAN WATER AND OF THE AIR OVER THE LAND.

Latitude,	Equator.	10°	20°	30°	40°	50°
Oceans,	2·3°	2·4°	3·6°	5·9°	7·5°	5·6°
Continents,	—	3·3	7·2	10·2	14·0	(25·4[1])

The ranges are more extreme in the marginal portions of the oceans, especially in those cases where there are seasonal alternations of warm and cold currents, as in the neighbourhood of Newfoundland and off the coast of eastern Asia. Observations of the temperature of the surface water in the Kattegat during the years 1880-1887, gave a mean of 1·6° in February, and 17·1° in July; which shows a range of 15·5°. At a depth of 26 meters, the minimum, in March, was 3·8°; the maximum, in October, was 11·9°. This gives a range of 8·1°, which is but little more than half as great as that on the surface, although the waters in the Kattegat must have been well mixed by the currents and the winds. In the Adriatic Sea, near Lesina, the following mean temperatures were obtained as the result of several years' observations :—

WATER AND AIR TEMPERATURES AT LESINA.

	Winter.	Spring.	Summer.	Autumn.	Year.	Range.
Air,	9·2°	14·8°	24·4°	17·9°	16·6°	15·2°
Water surface,	13·5	15·0	22·0	19·5	17·5	8·5
Water at depth of 10 meters,	13·9	14·7	20·3	18·4	16·8	6·4

The temperature at some depth below the surface of lakes and oceans affects the air temperature through the action of convectional currents, and the mixture of the water by means of winds and waves. These examples suffice to illustrate the conditions of diurnal and annual warming of bodies of water, and the effect of these conditions in lessening the extremes of air temperature.

Effect of clouds upon the temperature of oceans and continents.—The cooling of large bodies of water during the winter months is retarded

[1] North latitude only.

by the rise of the warmer strata from below, in consequence of the cooling of the surface water. There is, however, a second check which results from the difference in the conditions of the air over large surfaces of land and water; a difference which is of great importance because of its effect upon the amount of heat received through insolation and also upon that lost by radiation. Both of these values are diminished, the latter probably more than the former. The control in question is especially effective in the higher latitudes. The following data give some idea of the effect of a clear sky, and of a cloudy sky, upon the temperature of the air. These results were obtained by Kaemtz for Dorpat, and show, in degrees, how much the temperature at that station departs from the normal under different conditions of cloudiness.

EFFECT OF CLOUDINESS UPON AIR TEMPERATURE AT DORPAT.

Cloudiness (0—4)	0 clear.	1	2	3	4 overcast.
Winter, - - -	− 10·5°	− 6·8°	− 3·1°	+ 0·5°	+ 4·4°
Summer, - - -	+ 1·6	+ 0·8	− 0·3	− 1·2	− 2·7
Year, - - - -	− 3·7	− 1·9	− 1·0	− 0·2	+ 1·3

At latitude 58° N., a perfectly clear sky lowers the temperature more than 10° below the mean. When the sky is completely overcast, the temperature is 4·5° above the mean, because radiation is then almost completely checked. In summer, the effect is just the reverse, but it is also weaker. In the mean for the year, therefore, in the higher latitudes, the effect of the winter radiation predominates. Similar results have been obtained for other stations. In the case of Prague, Augustin found the winter temperature − 6·5° on perfectly clear days, and on overcast days, − 0·1°. The summer temperature at Prague, on clear days, is 21·0°, on overcast days, 16·1°; the mean for the year being 8·1° for clear and 8·0° for overcast days. In the case of Vienna, Kostlivy finds the temperature of a clear January, − 4·1°, of an overcast January, − 1·9°, a clear July, 21·7°, an overcast July, 17·6°, and an annual mean of 9·4°, clear; 7·6°, overcast. At Greenwich, the temperature on perfectly clear days in the winter is 3·0° below the mean, and normal on overcast days.[1] In summer, it is 2·5° above the mean on clear days, and 1·6° below it on overcast days. The annual mean is 0·6° higher for clear days than for overcast days. We may, therefore, conclude that a low percentage of cloudiness in

[1] In Greenwich, overcast days are the normal condition in winter.

higher latitudes means a considerable reduction of the winter temperatures, and a slight increase in the temperatures of summer, the result being a reduction of the mean annual temperature. In low latitudes, a decrease in cloudiness produces a decided increase in the mean annual temperature, which corresponds to the effect in the summer of higher latitudes. Furthermore, the sky over the oceans is more frequently cloudy, and the amount of cloudiness is also greater than it is over the continents, over whose interiors, in the case of the larger land masses, the cloudiness shows a marked decrease. Thus the various factors already referred to all combine to increase the contrasts of temperature between winter and summer over the lands, and to diminish them over the oceans. The resulting effect upon the mean annual temperature, which cannot easily be directly deduced for middle and higher latitudes, appears at a glance upon a chart of mean annual isotherms. These isotherms show a decrease in the mean annual temperature over the continental masses in higher latitudes, and an increase in this temperature over the continental areas of lower latitudes. The transition from one condition to the other takes place at about latitude 40°, where land and water have the same mean annual temperature.[1]

Mean temperatures in continental and marine climates.—The following table will serve to show more clearly the effect of the land upon the mean temperatures, and upon the annual range of temperature.

MEAN TEMPERATURES ALONG LATITUDE 52° N. FROM WEST TO EAST.[2]

Station	N. Lat.	Longitude.	Year.	Jan.	July.	Difference.
Valentia, - -	51° 54′	10° 25′ W.	10·1°	5·7°	15·1°	9·4°
Oxford, - -	51° 46′	1° 16′ W.	9·4	3·6	16·2	12·6
Münster, - -	51° 58′	7° 38′ E.	9·1	1·3	17·3	16·0
Posen, - -	52° 25′	17° 5′ E.	7·8	− 2·7	18·3	21·0
Warsaw, - -	52° 13′	21° 2′ E.	7·3	− 4·3	18·7	23·0
Kursk, - -	51° 45′	36° 8′ E.	5·7	− 9·4	19·8	29·2
Orenburg, - -	51° 46′	55° 7′ E.	3·3	− 15·3	21·6	36·9
Barnaul and Semipalatinsk	51° 52′	80° 30′ E.	1·7	− 18·0	21·8	39·8

[1] See the charts of the January, July, and mean annual isotherms in Bartholomew's *Physical Atlas*, Vol. III., Meteorology, Edinburgh, 1899.

[2] All temperatures have been reduced to a mean altitude of 100 m., and are therefore directly comparable.

Over a distance of 91° of longitude, namely from Valentia to Barnaul, the mean annual temperature decreases 8·4°, and the January temperature decreases 23·7°. The July temperature, on the contrary, increases 6·7°, and the mean annual range increases 30·4°. In other words, the range is quadrupled in going inland. Kursk is situated about half-way between Valentia and Barnaul-Semipalatinsk; thus it is seen that the changes are greater over the first half of the distance from Valentia to Kursk, than over the section from Kursk to Barnaul-Semipalatinsk, in which region the climate is purely continental.

A similar and more comprehensive summary, prepared by the author, showed that, between latitudes 47° and 52° north, on the continent of Europe, the temperature changes from west to east are as follows: In every 10° of longitude towards the east, there is a decrease of 3·1° in winter; an increase of 0·7° in summer, and a decrease of 1·3° in the mean annual temperature.

The lowering of the temperature produced by the continent in winter is thus seen to be more than four times as great in these latitudes as the raising of the temperature in summer. The net result must therefore be a reduction of the mean annual temperature. It may be assumed, *a priori*, that the warming effect will be the dominant one in low latitudes, where there is no actual winter; and the course of the isotherms confirms this conclusion. Nevertheless, a few illustrations of these conditions will not be out of place here. Taking first the case of eastern Australia, in latitude 31° S., we find the following: Port Macquarie (lat. 31·4°) on the coast, mean annual temperature, 17·6°; Murrurundi and Goonoo (lat. 31·5°), 161 km. inland, 18·4°; Cowga (lat. 31·2°) 470 km. inland, 21·6°.[1] For India, latitude 20°-21° N., the conditions are shown in the following table:—

INFLUENCE OF DISTANCE FROM THE SEA COAST UPON TEMPERATURES IN INDIA.

Station.	Latitude.	Longitude.	Year.	January.	July.	Difference.
False Point, -	20·3° N.	86·8° E.	26·7°	20·9°	30·1°	9·2°
Cuttack, - -	20·5 N.	85·9 E.	27·0	21·3	31·5	10·2
Nagpur, - -	21·1 N.	79·2 E.	27·6	21·5	35·4[2]	13·9

Between latitudes 15° and 16° N. we have: Goa (15·4° N.; 73·9° E.), on the west coast, with a mean annual temperature of 27·7°;

[1] All temperatures reduced to sea-level. [2] May.

Bellary (15·2° N.; 76·9° E.), in the interior, with 29·5°; and Masulipatam (16·2° N.; 81·2° E.), on the east coast, with 27·6°.

North of latitude 30°, land areas distinctly tend to raise the temperature; but this effect is proportionally less marked than the tendency to lower the temperature which appears in higher latitudes. Furthermore, it is not to be expected that the effect of the land in raising the temperature should be at a maximum near the equator. As a result of the uniformity of temperature throughout the year, and of the greater cloudiness and rainfall in the equatorial belt, higher mean temperatures are found north and south of the equator than on the equator itself. Indeed, when the land within the equatorial zone is covered with extensive forests, as is the case in the Amazon valley and in the Soudan, the mean temperature at a great distance inland is relatively low. It is lower, in fact, than over the oceans along the same parallels of latitude. The mean annual temperature is 25°–26° in South America, and is the same on the Congo, Equatorville, for example, having 24·3°, which, when reduced to sea-level, gives 25·9°. It is a very noteworthy fact that, near the equator, water can be warmed at least as much as the forested continents. The cooling action of radiation and of evaporation from forests wet by heavy rains is greater than that of the waters of the oceans. The greatest excess of temperature on lands as compared with that on oceans is found in the neighbourhood of the tropics.

Annual range of temperature in continental and marine climates.—A common feature of continental climates in all latitudes is their large range of temperature. Marine climates, on the other hand, are characterised by a small annual range. Stations which are freely exposed to the influence of winds from the ocean have an extraordinary uniformity of temperature. On the island of Monach (lat. 57° 32′ N.), one of the westernmost outposts of Europe, in the Atlantic Ocean, the temperature of the coldest month is 5·2°; and of the warmest month, 13·1°; which gives an annual range of 7·9°. Inverness, in the same latitude, on the eastern coast of Scotland, has a temperature of 3·2° in the coldest month; 14·2° in the warmest month; and a mean annual range of 11·0°. Port Stanley, in the Falkland Islands (lat. 51° 41′ S.), has a January mean of 9·8°; a July mean of 2·5°; and a mean annual range of 7·3°. On Kerguelen Island (lat. 49° S.), the difference between the temperature of winter and summer is but 5°; and even the minimum temperatures of winter are but little lower than those of summer. The greatest difference

on the same parallel of latitude (62° N.) is probably that between Thorshaven, on the Faroe Islands, and Yakutsk, in Siberia. Thorshaven has a mean temperature of 3·0° in March, and 10·9° in July; which gives a range of 7·9°. Yakutsk has a January mean of −42·8°; a July mean of 18·8°, with a range of 61·6°. In January, the difference in temperature between the Faroe Islands, in the North Atlantic Ocean, and the continental station in eastern Siberia, is 46·2°. Some examples of annual ranges of temperature in North America and in the British Isles have been given on page 11. A continental climate is, therefore, with good reason, considered to be *severe* as regards its temperature conditions, while a marine climate is *mild* by contrast. The geographical distribution of the *severe* and the *mild* climates can be seen upon the charts of annual ranges of temperature. Supan has drawn a chart of this sort to show the differences in temperature between the extreme months; and van Bebber has constructed one showing the difference between the mean absolute annual extremes.[1] In 1894, Connolly published a new chart of equal annual ranges of temperature, based on the "Challenger" isothermal charts.[2]

Annual march of temperature in continental and marine climates.—Continental climates differ from marine climates not only in the amount of their annual ranges of temperature, but also in having a different type of annual temperature curve. In continental climates, the maximum temperature comes about one month after the date of the sun's maximum altitude, while the minimum temperature is similarly, though to a less marked degree, delayed after the sun's lowest altitude. The only exception to this rule is that of the tropical monsoon climates, which have their highest temperature before the rains, and hence before the time at which the sun attains its maximum altitude. Examples of such conditions are furnished by India and Senegambia. In marine climates, on the other hand, the delay in the time of maxima and minima is much greater. The lowest temperature does not occur until two, or even three, months after the greatest declination of the sun, *i.e.*, in February or March. The highest temperature is likewise delayed, after the sun's greatest altitude, although to a less degree, the warmest month being August. Thus the temperature rises, as a whole, more rapidly than it falls. The characteristics of the annual march of temperature

[1] Bartholomew's new *Physical Atlas*, III., Meteorology, Plate 2.

[2] J. L. S. Connolly: "A New Chart of Equal Annual Ranges of Temperature," *Amer. Met. Jour.*, X., 1893-94, 505-506. This chart is reproduced in Bartholomew's *Atlas of Meteorology*, Plate 2.

in a marine climate are a cold spring and a warm autumn, April and May being colder than October and September. The conditions are reversed in continental climates, in which the temperature rises more rapidly, April being warmer than October, as a general rule, provided there is no snow on the ground.

Graphic and tabular illustrations of the annual march of temperature in continental and in marine climates are given in the table on page 142, and in Fig. 3. In a marine climate there is hardly any difference in the annual march of temperature at latitudes 35° and 60°. The only difference which does exist is that the spring is somewhat colder, and the autumn warmer, at latitude 35°. In a continental climate, on the

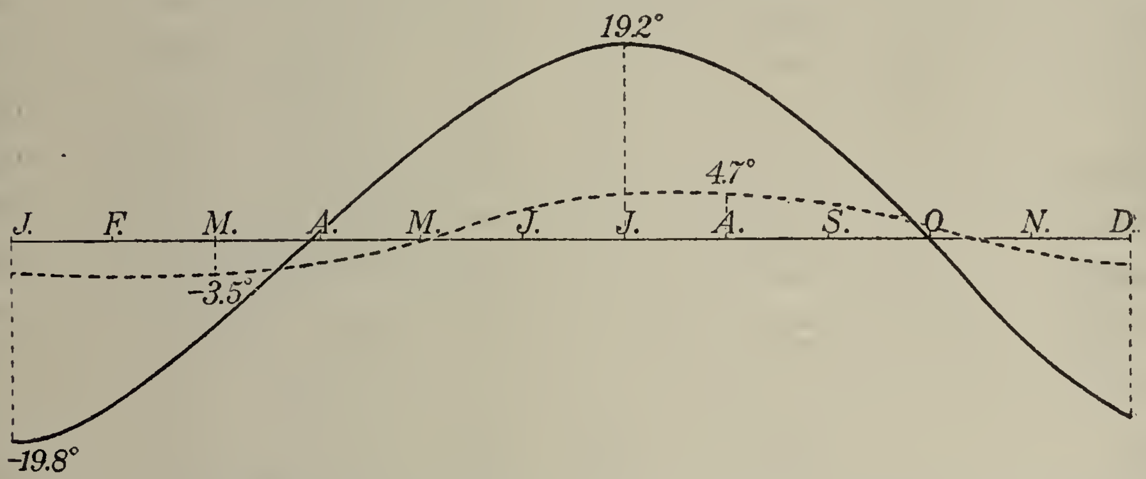

Fig. 3.—Annual March of Temperature in Continental and Marine Climates.

other hand, there is a very great difference in the amount of the annual range of temperature. At the same time, the spring is warmer, and the autumn relatively colder, at latitude 40° than at latitude 60°.

Influence of continental and marine climates upon crops.—According to Schindler, continental and marine climates have different effects upon the size of the crops of cereals, and also upon the relative amounts of the most important elements of nutrition which these cereals contain. In a marine climate, wheat contains only 9-12 per cent. of protein; and for this reason, the consumption of meat, leguminous plants, and other nitrogenous foods must be increased. In a continental climate, such as that of southern Russia and Hungary, where the time of growth is shorter, wheat is 4-8 per cent. richer in protein; and the need of other nitrogenous food-stuffs is lessened. A hot, dry climate decreases the proportion of starch, and increases that of gluten.[1]

[1] Schindler: *Der Weizen in seinen Beziehungen zum Klima*, Berlin 1893. See also Wollny: *Forsch. a. d. Geb. d. Agrikulturphysik*, VIII., 1885, 313; XVII., 1894, 209. Reference will later be made to other relations of climate and vegetation.

ANNUAL MARCH OF TEMPERATURE IN A MARINE AND IN A CONTINENTAL CLIMATE.[1]

DEPARTURES OF THE MONTHLY MEANS FROM THE ANNUAL MEAN.

Latitude.	Marine Climate.		Continental Climate.		Marine Climate.	Continental Climate.
	35°	60°	40°	60°	Mean.	Mean.
January,	-3·4°	-3·2°	-15·3°	-24·4°	-3·3°	-19·8°
February,	-3·5	-3·4	-13·9	-19·5	-3·4	-16·7
March,	-3·5	-3·5	-5·6	-10·0	-3·5	-7·8
April,	-2·2	-1·6	1·7	0·7	-1·9	1·2
May,	-0·5	0·3	8·0	10·7	-0·1	9·4
June,	2·0	3·2	12·2	20·6	2·6	16·4
July,	4·2	4·4	14·2	24·2	4·3	19·2
August,	4·8	4·7	12·4	20·0	4·7	16·2
September,	3·9	3·2	6·6	12·5	3·6	9·5
October,	1·6	0·7	-0·8	0·9	1·2	0·1
November,	-0·4	-2·0	-7·1	-13·8	-1·2	-10·4
December,	-2·8	-2·8	-12·2	-22·3	-2·8	-17·3
Year,	18·4	7·6	15·8	-4·8	13·0	5·5
Range,	8·3	8·2	29·5	48·6	8·2	39·0

The influence of a snow surface on temperature is a subject whose great importance Woeikof first pointed out.[2] Viewing the matter from a climatic standpoint, we may say that the winter cold of high latitudes changes oceans and lakes into land masses; for these bodies of water are then covered with a sheet of ice, which temporarily prevents the water surface from having any influence upon the temperature of the air. In a similar way, the winter snow covering over the continents alters the effect which the land has upon the

[1] Marine Climate—I. Bermuda, Madeira, Azores; four stations. Mean latitude, 35°·3 N., 33°·7 W. II. Hebrides, Orkney, Shetland, and Faroe Islands; five stations; 60°·0 N., 3°·9 W. Continental Climate—I. Turkestan; twelve stations between Merv and Nukus, Kisyl Arwat and Yarkand; 40°·0 N., 66°·5 E., altitude, 520 meters. II. Siberia; nine stations between Tobolsk and Yakutsk, together with ½ (Enisseysk, Yakutsk). Mean latitude, 60°·3 N.; longitude, 104°·4 E., altitude, 200 m.

[2] A. Woeikof: "Der Einfluss einer Schneedecke auf Boden, Klima und Wetter," Penck's *Geograph. Abhandl.*, III., No. 3, Vienna, Hölzel, 1889 [*M.Z.*, VI., 1889, (65)—(68)]. Brückner has recently discussed the influence of a snow cover upon the climate of the Alps. See *Das Wetter*, XVII., 1900, 193-208, 222-234.

temperature of the air. In the first place, the snow prevents the ground from having any control over the temperature of the air, and substitutes its own influence for that of the ground. The influence of the snow cover during the season of frost is to increase the radiation, and at the same time to prevent the flow of heat from the ground by conduction.[1] For this reason, radiation from a snow surface under the clear sky of a continental winter considerably reduces the winter temperatures and increases the annual range of temperature. At the same time, the snow cover does protect the ground to a marked degree against the penetration of frost, and in this respect it is very beneficial. On the other hand, when the temperature begins to rise in the spring, a snow cover effectively retards the warming of the air, as do the frozen surfaces of oceans and lakes, for the reason that the heat of the sunshine and of the warmer air currents is almost wholly expended in melting the snow and ice.

It is a mistake to suppose that snow cannot melt in the sun until the temperature of the air rises above the freezing point, and that, as long as the temperature is below 0°, snow can be melted only by warmer air coming from regions where there is no snow on the ground.[2] Snow is known to melt in the sun even when the air temperature is below freezing, and this fact is readily explained according to physical laws. In Spitzbergen, in the middle of May, Ekholm found the mean temperature of the snow surface 1° higher than the maximum air temperature. In a snow layer 2 meters thick, the thermometer showed a rapid warming from – 7° to 0°, although the mean air temperature during this time was – 6°, and the maximum, – 1·1°. The snow melted rapidly, although the air temperature did not rise to 1° – 2° above freezing until five days later.[3]

The increase of temperature in the spring is much retarded by the presence of snow, the retardation being greater the thicker the snow cover. Hence it follows that the spring is colder than the autumn, April being colder than October; and the annual march of temperature approaches that in a marine climate. In cases where the winter precipitation in high latitudes is small, the temperature rises very rapidly over the dry land in the spring, and the increase of temperature in the air more closely follows the sun. April is then warmer than October. The differences in the annual march of temperature in continental

[1] See L. Satke: "Fünfjährige Beobachtungen der Temperatur der Schneedeck in Tarnopol," *M.Z.*, XVI., 1899, 97-106.

[2] A. Woeikof: *Die Klimate der Erde*, I., 71-72.

[3] J. Hann: "Die Ergebnisse der schwedischen internationalen Polarexpedition, 1882-83, auf Spitzbergen, Kap Thordsen," *M.Z.*, XI., 1894, 41-53.

regions which have no snow cover, and in those which are deeply covered with snow, have been clearly pointed out by Woeikof.[1]

In Turkestan, Central Asia, and Mongolia, where there is little snow, October is colder than April, and November is colder than March. In the very same latitude, on the high plateau of Armenia, where there is a heavy snowfall, the conditions are reversed. For example, Erivan, Kars, and Alexandropol, in latitude 40·5° N., have an annual range of 32°; October is 2·6° warmer than April, and November is 2·9° warmer than March. On the other hand, Nukus, Taschkent, and Kaschkar, in latitude 41·1° N., have an annual range of 31·5°; October is 4·2° colder than April; and November is 3·5° colder than March. In both regions the climate is continental, but the annual march of temperature differs very much in the two cases. In general, April is colder than October in middle and higher latitudes, because of the winter snow cover which is almost always present. In higher latitudes, May is also colder than September, even in a continental climate.

The diurnal range of temperature in continental and marine climates: the effect of water vapour.—The diurnal range of temperature increases with increasing distance from the ocean, as does the annual range. The greater and more rapid warming of the land surface, and the stronger insolation, increase the daily temperature maxima; while on the other hand the clear sky and dry air at night favour rapid radiation and cooling of the earth's surface, and produce low nocturnal minima. Thus there result a large diurnal and a large annual range.[2]

Tyndall, more than any one else, has maintained that with a larger quantity of water vapour in the atmosphere, radiation is checked, and this diminishes the diurnal and annual ranges of temperature in marine and in littoral climates. He has also tried to prove this by means of observations.[3] Although Tyndall probably over-estimated this effect of the water vapour, recent observations of the absorption of certain infra-red waves leave no doubt whatever of the fact that such an influence really exists. Strachey has obtained some notable results from certain observations made in Madras.[4]

[1] A. Woeikof: "Kontinentales und Ozeanisches Klima," *M.Z.*, XI., 1894, 1-9.

[2] For examples from the United States see A. M'Adie: "Mean Temperatures and their Corrections in the United States," *U.S. Signal Office*, Washington, D.C., 1891, pp. III. and IV.

[3] John Tyndall: "Note on Terrestrial Radiation," *Nature*, XXVII., 1882-83, 377-379 (*Z. f. M.*, XVIII., 1883, 274-275).

[4] "On the Action of Aqueous Vapour on Terrestrial Radiation," *Philosoph. Mag.*, London, XXXII., 4th Ser., 1866, 64-68. See also F. W. Very: "Atmospheric Radiation," *U.S. Department of Agriculture, Weather Bureau, Bulletin G*, 4to, Washington, D.C., 1900.

Observations made between March 4 and 25, under a perfectly clear sky, showed the following decrease of temperature by night, between 6.30 P.M. and 6.30 A.M. :—

EFFECT OF VAPOUR PRESSURE UPON AIR TEMPERATURE AT MADRAS.

Vapour Pressure, - -	22·4	18·0	15·4	11·0 mm.
Change of Temperature, -	3·7°	5·7°	6·7°	9·2°

The more water vapour the air contains, the more readily do light fogs form, although these may be invisible to the eye. It is therefore still uncertain whether the effect of the vapour pressure here shown is to be attributed altogether to the vapour contents of the air. Woeikof believes that the connection between radiation and relative humidity is much more clearly proved.[1] Sutton has recently attempted to show the effect of the water vapour in the air upon nocturnal cooling by means of some observations at Kimberley, South Africa. For this purpose, he selected the observations made on perfectly clear nights during an entire year. No relation between the dew-point at 8 P.M. and the amount of nocturnal cooling could be established; but that there was a marked relation between the relative humidity and the nocturnal cooling is shown by the following figures :—

RELATIVE HUMIDITY AND NOCTURNAL TEMPERATURES AT KIMBERLEY.

Relative Humidity at 8 p.m. (Percentages).				
25-39	40-49	50-59	60-69	70-80 and above.
Mean Difference of Temperature between 8 p.m. and the Minimum.				
10·3°	8·8°	7·6°	7·2°	6·0°
(42)	(30)	(37)	(45)	(18)[2]

In the light of these data, there can no longer be any doubt that an increasing amount of water vapour in the air diminishes the extremes of temperature.[3]

[1] "Terrestrial Radiation and Professor Tyndall's Observations," *Nature*, XXVII., 1882-83, 460-461 (*Z. f. M.*, XVIII., 1883, 275-276).

[2] Number of observations.

[3] J. A. Sutton: "Aqueous Vapour and Temperature," *Symons' Met. Mag.*, XXX., 1895, 104-107.

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At Lisbon, on the coast of Portugal, the greatest mean diurnal range of temperature (August) is 6·6°, while at Madrid, at the centre of the peninsula, it is 14·5° in July. On the island Lesina (43·2° N.), the maximum (July) is 4·9°; at Tiflis (40·7° N.), 10·3°, and at Nukus, on the Sea of Aral (42·5° N.), it reaches 16·4°. At Kurrachee, on the coast of the arid Sind (lat. 24° 47′ N.), the greatest mean diurnal range of temperature is 14° in November,[1] and at Deesa (lat. 24° 16′ N.), on a dry, sandy plain in the interior, it is 18·7° in November.[2] The mean diurnal range of temperature in August (10 years) is 6·1° at San Diego, Cal., and 15·3° at Yuma, Ariz.

The diurnal range is greatest on deserts, and especially on dry plateaus. In Death Valley, California, the mean diurnal range of temperature in August, 1891, was 17·8°, and the greatest daily range in the same month was 23·3°. Rohlfs found the average difference in temperature between sunrise and 3 P.M. at Murzuk as high as 15·5° in winter. In the Kufra oasis, in latitude about 25° N., and at an altitude of 500 m. above sea-level, the mean difference for one month (August 15—September 14) was 22·2°, and at Audjila (lat. 29° N.) it was 19·7° in May. Similar results were obtained by Nachtigal, during the summer, in the desert between Murzuk and Kuka, the average diurnal ranges being between 19° and 22°. The diurnal ranges of temperature are probably the greatest in the interior of South Africa and of Australia. According to Livingstone, the average difference of temperature between sunrise and noon in the interior of Africa was 26·6° in June. The difference of temperature between sunrise and afternoon, in deserts, amounts in individual cases to 30°–40°, as the result of nocturnal radiation and of the general heating of the surface during the afternoon. The temperature of the sand reaches 70°–80°. Pechuël Loesche frequently observed surface temperatures as high as 80° on the Loango coast; and once, in February, noted a temperature of 84·6°. On December 25, 1878, at Bir Milrha, which is south of Tripoli, and 314 m. above sea-level, Rohlfs noted a thermometer reading of −0·5° in the morning, while in the afternoon the temperature was 37·2°, a difference of 37·7°. On the morning of May 25, 1840, Perrier found the ground around his tent in the Algerian Sahara covered with frost, while at 2 P.M. the thermometer stood at 31·5° in the shade. In northwestern Australia, Mitchell observed −11·6° at sunrise on the morning of June 2, and at 4 P.M. the reading was 19·4°. Livingstone and Wetzstein report, in the cases of South Africa and of Hauran, respec-

[1] The annual range is 9·4°. [2] The annual range is 14·8°.

tively, that the rocks which have been heated during the day occasionally cool so rapidly after sunset that they break into pieces with loud reports. Indeed, it seems probable that this rapid change of temperature is one of the agencies by means of which the rocks of the deserts are gradually disintegrated, notwithstanding the lack of rainfall, and are thus changed into fine material which may then be blown about by the winds.[1]

What a striking contrast there is between these great ranges and the temperature of the air over the open oceans! An immense number of observations taken on board ship in the tropical Atlantic Ocean between the equator and latitude 10° N., gave a diurnal range of only 1·6°, and a monthly range of only 6·5°. The results of the "Challenger" Expedition show a daily range of temperature over the open ocean between the equator and latitude 40° of 1·3° to 1·7°, while the water temperature varies only from 0·4° to 0·5°.[2]

The variability of the monthly means of temperature is greater in a continental than in a littoral or an insular climate. Hence a longer period of observation is necessary in the former than in the latter in order to give mean temperatures within the same degree of accuracy. The following table gives a few examples of this. The figures show the average departures of the means of single months from the general means of those months.

VARIABILITY OF MONTHLY MEANS OF TEMPERATURE.

Siberia and Ural, - -	Maximum, Dec., 3°·1	Minimum, July, 1°·2
Russia—Interior, - -	,, ,, 3°·5	,, May, 1°·4
Europe—West, - -	,, Jan., 2°·3	,, Sept., 1°·1
England, - - - -	,, ,, 1°·5	,, ,, 0°·9

The annual means, derived from the average departures for the months, amount to 2·0° for the first two groups, and for the last two, 1·4° and 1·2°. The mean departures are also very considerable in the interior of North America. Thus in February, the mean departure is 2·6°; in August, 1·1°, and for the year, 1·7°.

[1] See J. Walter: *Die Denudation in der Wüste* (Leipzig); and "Ueber Ergebnisse einer Forschungsreise auf der Sinaihalbinsel und in der Arabischen Wüste," *Verhandl. Berlin. Gesell. Erdkunde*, XV., 1888, 244-255. Walter believes that the daily range in temperature on the upper surface of dark rocks may be set as high as 80° as a result of insolation by day and radiation by night. The extremes may reach 41° and 3°, even in the shade.

[2] A. Buchan: "Challenger Reports," *Physics and Chemistry*, Vol. II., Part V., 1889.

It may therefore be stated as a general rule that the temperatures in a littoral or an insular climate are characterised by a greater uniformity, *i.e.*, by smaller variations about the mean, than in a continental climate. This is due, in the first place, to the influence of neighbouring large bodies of water, whose temperature changes are slow and gradual, and do not exceed certain narrow limits. Secondly, there is the moist atmosphere, the effect of which is to minimise the influence of cooling agencies by causing a condensation of water vapour and thus, by means of the latent heat liberated, decidedly diminishing the amount of cooling which would otherwise take place.

CHAPTER VIII.

INFLUENCE OF CONTINENTS UPON HUMIDITY, CLOUDINESS AND PRECIPITATION.

Decrease of humidity toward continental interiors.—The water vapour in the atmosphere is chiefly supplied by the evaporation of the ocean waters. Hence the amount of water vapour in the air naturally decreases with increasing distance from its main source of supply, becoming less and less toward the continental interiors. The amount of water vapour in the air of the interior of the larger continents, and even of deserts, is nevertheless greater than is commonly supposed. There are, furthermore, secondary sources of supply of atmospheric humidity, namely, evaporation from moist soil,[1] from rivers and lakes, and from vegetation, which must not be lost sight of.

Absolute and relative humidity of continental interiors in summer.—The following figures give some idea of the vapour contents of the air in the central portion of our largest continent in summer. They are the July means of absolute humidity in the steppes and the deserts of southwestern Siberia and western Turkestan, the vapour pressures being expressed in millimeters—Orenburg, 11·6; Uralsk, 11·2; Kasalinsk, 10·7; Aralsk, 10·9; Barnaul, 11·1; Nukus, 13·1; Petro-Alexandrovsk, 9·4; in Turkestan, Tashkent, and Margelan, 11·0. Even Yarkand, in Eastern Turkestan, has 12·3 mm. of vapour pressure in July. There is thus an average amount of water vapour equivalent to a pressure of 11 mm., which corresponds to the vapour contents of the

[1] The meaning of this statement should not be misunderstood. If it be assumed that the winds from the ocean deposit all their water vapour in the form of rain at a short distance from the coast, then this coastal strip will in its turn supply water vapour for the country farther inland.

air of Vienna, or even of Paris, in July, and is not much less than that in the same latitude on the west coast of Europe.

According to observations made by Rohlfs, the vapour contents of the air in Ghadames, in July and August, were equal to a pressure of 9·8 and 11·9 mm., and in the oasis of Kauar, in the heart of the Sahara (lat. 18°·8 N.), the absolute humidity in May was 13·0 mm. In the Lybian Desert, at the oasis of Kufra (lat. 24°·5 N.), the mean vapour pressure in the second half of August was 8·3 mm., and during the first half of September, 10·1 mm. This is the same as the average absolute humidity of the air in Vienna in May and June. Even in the most extreme cases, the vapour pressure was always at least 5—6 mm. Lichtenberg was therefore not very far from the truth when he said: "If we could produce cold as easily as we can light a fire, we could readily obtain water from the atmosphere, even in the desert." According to observations made by Rohlfs, the air at the Kufra oasis in August would have to be cooled 21·5° on the average, or from 30·0° to 8·5°, in order to bring about a condensation of the water vapour in it. At 3 P.M. on August 14, the dew-point was actually 39° below the air temperature, which was 38·9°. Although the absolute atmospheric humidity in the continental interiors during the heat of summer is by no means inconsiderable, the air is nevertheless far from being saturated. The relative humidity is very low, and the evaporating power of the climate is very great.

In regions of summer rains, the relative humidity even far inland is as high as on the coasts. Woeikof cites the following example. At Dorpat, the summer mean absolute and mean relative humidity are 10 mm. and 73 per cent., respectively; and at Yeniseisk, in Central Siberia, 9·0 mm. and 70 per cent. In July, the vapour pressure at Dorpat is 10·9 mm., and at Yeniseisk, 10·0 mm. The relative humidities are 73 per cent. and 69 per cent. The mean relative humidity for the stations in southwestern Siberia (excepting Barnaul) and in western Turkestan, named in the first paragraph of this section, is 45-50 per cent. in July, while it hardly falls below 75 per cent. on the western coast of Europe at the same time. During the driest months, the percentages are still lower. At Nukus, the relative humidity in June is 46 per cent., and for 2 P.M. it is only 19 per cent. Petro-Alexandrovsk, a degree and a half of longitude to the east of Nukus, in the desert, has 34 per cent. in June; Aralsk and Kasalinsk have 45 per cent. in July. The dryness is still more excessive on the steppes and deserts of lower latitudes. Thus

Ghadames has 27 per cent. in July, and 33 per cent. in August; the oasis of Kauar has 28 per cent. in June; the oasis of Kufra, 27 per cent. in August; 33 per cent. in September; and 17 per cent. for the 3 P.M. August mean. In the Punjab and northwestern provinces of India, the mean relative humidity in May is 31 per cent. at Lahore, 36 per cent. at Agra, and it is 30 per cent. at Ihansi in April. The lowest mean annual relative humidities in the United States are those for the Weather Bureau stations in the dry southwest. Yuma, Arizona, has a mean annual relative humidity of 42·9 per cent., with a mean monthly minimum of 34·7 per cent. in June. Santa Fé, New Mexico, has a mean annual of 44·8 per cent., with a mean monthly minimum of 28·7 per cent. in June. Pueblo, Colorado, has a mean annual of 46·2 per cent., with a mean monthly minimum of 37·6 per cent. in April. Death Valley, California, was found to have a mean relative humidity of 23 per cent. during five months (May-September) of the year 1891, when a temporary meteorological station was maintained there by the United States Weather Bureau.

Absolute and relative humidity of continental interiors in winter.—The atmosphere is very dry, *absolutely*, over the continents of middle and higher latitudes in winter, on account of the severe cold which then prevails, but it is *relatively* very moist and near saturation. Under these conditions, the relative humidity does not decrease inland, but it actually increases. The January mean for the stations in southwestern Siberia and western Turkestan is 1·7 mm. for vapour pressure, and 86 per cent. for relative humidity. Yarkand has means of 1·3 mm. and 58 per cent. In low latitudes, where the temperatures are high, the air is relatively dry in the interior, even in winter. Thus Rohlfs' observations showed a winter mean, at Murzuk (lat. 25° 54′ N.), of 4·6 mm. for vapour pressure, and 47 per cent. for relative humidity.

Evaporation.—A number of measurements of evaporation in the Russian Empire are available, which, in view of the similar exposure of similar apparatus at all the stations, furnish fairly comparable relative values of this climatic element. For this reason, some of the most important results of these observations are given here.[1]

[1] E. Stelling: "Ueber den jährlichen Gang der Verdunstung in Russland," *Repert. f. Met.*, VII., No. 6, 1880, and O. Britzke, on the same subject, *ibid.*, XVII., No. 10, 1894. Rainfall and evaporation relate to the same years. The amounts of evaporation are comparable with one another, but do not represent the evaporation from a free water surface in the sun, nor the evaporation from the ground. The latter naturally depends chiefly upon the amount of moisture in the ground.

EVAPORATION AND RAINFALL IN RUSSIA.

Station.	North Latitude.	East Long.	Rainfall (cm.)	Evaporation (cm.)
St. Petersburg, - - -	59°·9	30°·3	47	32
Moscow, - - - -	55°·8	37°·6	54	42
Kiev, - - - - -	50°·5	30°·5	53	48
Kishinev, - - - -	47°·0	28°·8	47	55
Lugan, - - - - -	48°·6	39°·3	37	74
Astrakhan, - - - -	46°·4	43°·0	16	74
Kasalinsk, - - - -	45°·8	62°·1	10	106
Nukus, - - - - -	42°·5	59°·6	7	193
Petro-Alexandrovsk, - -	41°·5	61°·1	6	232
Sultan Bend, - - - -	37°·0	62°·4	18	276
Tashkent, - - - -	41°·3	69°·3	33	134
Peking, - - - - -	39°·9	116°·5	62	91
Katherinburg, - - -	56°·8	60°·6	36	45
Barnaul, - - - -	53°·3	83°·8	26	57
Akmolinsk, - - - -	51°·2	71°·4	23	104

These data clearly show the *decrease* in the rainfall toward the interior and also from north to south, while the amount of the evaporation *increases* very rapidly in these directions.

Cloudiness.—The greater relative dryness of the continental interiors naturally involves also a smaller amount of cloudiness, especially in summer. While the mean cloudiness in northwestern Europe is 68 per cent., this amount decreases southeastward toward the interior. In Russia, the maximum cloudiness (70-75 per cent.) is in the northeast. From there the cloudiness decreases southward, toward the Black Sea, to 50-60 per cent. on the southern coast of the Crimea. Further southeast there is a still greater decrease, for on the Sea of Aral the amount is 40 per cent., and in western Turkestan it decreases to 35 per cent. The cloudiness is 70 per cent. on the White Sea in summer; in central Russia and western Siberia it is 50 per cent.; on the Caspian and the Sea of Aral it is 35 per cent. and 20 per cent.; and farther south, in western Turkestan, it is 10 per cent. The number of clear days during the year, which is but 20 on the peninsula of Kola and on the White Sea, rises to between 160 and 180 on the

steppes of western Turkestan.[1] In the United States, the maximum cloudiness is found on the extreme northwestern Pacific coast (65 per cent.). Over the Lake region, there is a considerable district with a mean annual cloudiness of 60 per cent. The greater part of the middle and south Atlantic coast has a percentage of 50, and the average cloudiness over most of the Mississippi and Missouri valleys is between 45 per cent. and 50 per cent. In southern Nevada, southeastern New Mexico, most of Arizona, and the southern and eastern sections of California, except on the immediate sea coast, the mean annual cloudiness is 30 per cent.[2]

The least cloudiness for the world as a whole is probably to be found in northern Africa, Arabia, the desert regions of Arizona and New Mexico, and perhaps also in the interior of Australia. Cairo has a mean cloudiness of 19 per cent. The chart of mean annual cloudiness[3] shows that in Europe the decrease in the mean cloudiness is mainly in the direction from north to south.

Precipitation.—The amount and frequency of precipitation as a rule decrease inland, but this decrease is so irregular, and depends so much upon the topography; upon the position of the mountain ranges with respect to the rain-bearing winds, etc., that no general illustrations of this rule can be given. The distribution of the rainfall over the continents is therefore best left for consideration in the study of the climatology of special districts.

[1] A. Schönrock: "Die Bewölkung des Russischen Reiches," *Mem. Acad. Imp. Sci. de St. Petersbourg*, VIII. Ser., *Cl. Phys.-Math.*, I., No. 9, 1895. [*M.Z.* XII., 1895, (89) – (92)].

[2] See map of Average Cloudiness in the United States in *Report of the Chief of the Weather Bureau*, 1896-97, Part VI., Chart XX., 4to, Washington, D.C., 1898.

[3] See Bartholomew's new *Atlas of Meteorology*, Plate 18.

CHAPTER IX.

INFLUENCE OF CONTINENTS UPON WINDS.

Land and sea breezes.—The influence which land areas have upon the general circulation of the atmosphere, or, in other words, upon the prevailing winds, is of the greatest climatic importance. This influence is due, as has already been seen, to the varying differences of temperature between land and water, from day to night, and especially from summer to winter. The distribution of temperature, humidity, cloudiness, and rainfall are all dependent upon the prevailing winds. We shall first consider a local phenomenon which results from these differences of temperature, and which is limited to the sea-coasts.

Of the periodic atmospheric movements which are produced by the contrast in temperature between land and water, land and sea breezes were the first to become known. With the change from day to night, the relative temperature conditions of land and water are reversed, and this reversal is followed by a corresponding change of winds. In low latitudes, where there is no real winter, these periodic winds blow throughout the year. In higher latitudes, they occur almost exclusively in the warmer months. When the sea breeze first begins to blow, its course is in general at right angles to the coast, but as the air is later drawn in from a greater distance off-shore, it comes under the influence of the deflective force of the earth's rotation, and is turned to the right in the northern hemisphere. Thus, if the sea breeze blows from the east at the start, it gradually turns toward the south during the day, and the land breeze which begins in the west turns to the north during the night. The origin of land and sea breezes was from the earliest times attributed to the unequal warming and cooling of land and water, but the physical explanation of these winds was not given until within recent years. The accuracy of the explanation has only lately

been established by a study of the difference in the variations of pressure on the coast and inland.

The phenomena of land and sea breezes, and of their changes, should be clearly in mind at the outset. "The inhabitants of the sea-shore in tropical countries," says Maury,[1] "wait every morning with impatience the coming of the sea breeze. It usually sets in about ten o'clock. Then the sultry heat of the oppressive morning is dissipated, and there is a delightful freshness in the air which seems to give new life to all for their daily labours. About sunset there is again another calm. The sea breeze is now done and in a short time the land breeze sets in. This alternation of the land and sea breeze—a wind from the sea by day and from the land by night—is so regular in tropical countries that it is looked for by the people with as much confidence as the rising and setting of the sun." The sea breeze is refreshing and health-giving, not only by reason of its relatively low temperature, but also because it brings in pure air from the ocean and dispels the miasmas which so frequently make low, tropical coasts below high tide level extremely unwholesome. In these regions, one of the most essential requisites of a wholesome location and of a wholesome climate is the free access of the sea breeze. The land breeze, on the other hand, is often distinctly unwholesome, especially on the tropical west coast of Africa; and it may give rise to actual epidemics if it blows for an unusually long period. This is especially the case where stagnant bodies of water, within reach of high tide, and surrounded by luxuriant vegetation, stretch inland from the coast.

When the sea breeze has the same direction as the prevailing winds, it often increases in the afternoon to the velocity of a gale, while the land breeze is hardly noticeable. "In the summer of the southern hemisphere," says Maury, "the sea breeze at Valparaiso is more powerfully developed than at any other place to which my services afloat have led me.[2] Here regularly in the afternoon, at this season, the sea breeze blows furiously; pebbles are torn up

[1] *The Physical Geography of the Sea*, 15th ed., London, 1874, Chap. VI., pp. 133-145.

[2] The cause of this unusual development probably lies in the combination of three circumstances, namely, the abnormally low temperature of the ocean waters; the excessive heating of the dry, barren land under the strong insolation of the southern summer; and a prevailing wind from the southwest which very nearly coincides with the direction of the sea breeze. Similar conditions, attended by similar consequences, are noted on the southwestern coast of Africa, and on the coast of California.

from the walks and whirled about the streets; people seek shelter; the Almendral is deserted, business interrupted, and all communication from the shipping to the shore is cut off. Suddenly the winds and the sea, as if they had again heard the voice of rebuke, are hushed and there is a calm. The lull that follows is delightful. The sky is without a cloud; the atmosphere is transparency itself; the Andes seem to draw near; the climate, always mild and soft, becomes now doubly sweet by the contrast. The evening invites abroad, and the population sally forth—the ladies in ball costume, for now there is not wind enough to disarrange the lightest curl. In the southern summer, this change takes place day after day with the utmost regularity, and yet the calm always seems to surprise, and to come before one has time to realise that the furious sea-wind could so soon be hushed."[1]

Height of the sea breeze.—No general statements can be made as to the height of the sea breeze, or as to its distance of penetration inland, for these features primarily depend upon local conditions. At Coney Island, which is near New York City, and is therefore a long distance outside of the tropics, some observations as to the height of the sea breeze in summer (August) were recently made by means of a captive balloon. The sea breeze extended above this low island to a height of about 150 meters, and the upper current, from the opposite direction, was encountered at about 200 meters. Balloon ascensions have frequently shown that the air blows seaward above the sea breeze, for the balloons have been driven out to sea by this upper, off-shore wind. By causing the balloon to descend into the lower current, it could again be brought back to the land. By means of this method of observation, the height of the sea breeze at Toulon, in the middle of October, 1893, was found to be about 400 meters, while a distinct off-shore current was found at 600 meters.

On the coast of southern California, the sea breeze blows throughout the greater part of the year. It is weak in winter and strong in summer. During the latter season, it is a dry wind even on the coast.[2] The height of this sea breeze may be assumed to be about $2\frac{1}{2}$ kilometers, as determined by the smoke

[1] M. F. Maury, *loc. cit.* Note also in this same book the graphic description given by Lieutenant Jansen of the weather conditions which accompany the changes of land and sea breezes on the coast of Java.

[2] The cause is undoubtedly to be found in the fact that the ocean near the coast is very cool, while on the other hand the land is very much warmed. The air which comes off the ocean must therefore seem relatively dry over the land.

of the brush fires upon the mountains. The movement of this smoke is toward the east at first; at a greater height, this drift becomes progressively less and less marked, until, at an altitude of about 2500 meters above the sea, there is a calm. The smoke which continues to rise above this level turns to the west. This observation can, however, be made only by climbing up on a mountain. On the summit of Old Gray Back, the wind may be felt moving westward, while 1800 meters lower the sea breeze is blowing in the canyons.[1] This sea breeze upon the coast of southern California is, however, a wind which partakes rather of the character of a monsoon, because it is an effect of the prevailingly higher temperature in the interior of California as compared with the ocean.

The off-shore beginning of the sea breeze was noted as long ago as the days of Dampier. He called attention to the fact that the sea breeze begins out at sea, and gradually extends to the shore, while, on the contrary, the land breeze begins at the shore and forces its way out to sea. The observer on the shore detects the approach of the sea breeze out at sea by the rippling of the surface, and by the deep blue colour which the water assumes, while near shore the water is still as smooth and shining as glass. In about half an hour, the sea breeze reaches the land, and increases in velocity till afternoon. The land breeze first reveals its presence out at sea by the odour of plants and flowers, which suddenly fills the air before the breeze itself is noticeable.

An increase of the sea breeze with a flood tide, when the flood tide and sea breeze occur simultaneously, has frequently been observed. Krümmel is probably right in explaining this fact as due to the raising of the atmospheric strata by the tide, and the resulting increase in the pressure gradient toward the land.[2]

The sea breeze in New England.—The most exhaustive study of the sea breeze in any special area was that undertaken by the New England Meteorological Society during the summer of 1887, when a series of systematic observations was carried out at 130 stations on the coast of Massachusetts.[3]

In this region, as elsewhere, the sea breeze begins first over the open ocean and gradually works in toward the coast, which it reaches at

[1] T. S. Van Dyke: *Southern California*, New York, 1886.

[2] *Geophysikalische Beobachtungen der Plankton-Expedition*, Leipzig, 1893, 44-50. Dampier's classical description of land and sea breezes is reprinted in *Amer. Met. Jour.*, IV., 1887-88, 59-61.

[3] W. M. Davis, L. G. Schultz, and R. De C. Ward: "An Investigation of the Sea Breeze," *Annals of the Astronomical Observatory of Harvard College*, Vol. XXI., Part II., Cambridge, Massachusetts, 1890. See also W. C. Appleton: "The Sea Breeze at Cohasset, Mass;" *Amer. Met. Journ.*, IX., 1892-93, 134-138.

eight o'clock on very warm, quiet mornings; but the hour is more often nine or ten o'clock, and not infrequently the arrival is delayed until noon. The sea breeze advances inland from the shore-line at a rate of from 5 to 13 km. an hour at first, but the rate afterwards becomes slower, when the breeze approaches its greatest penetration of 16 to 30 km. late in the afternoon. On sea breeze days, the average velocity of the wind is 23 km. an hour at 3 P.M., at Boston, the only coast station with an anemometer. This is much faster than its rate of penetration inland, and an ascent of the breeze at its inland margin must therefore be inferred. The development of the afternoon thunderstorms which accompany sea breezes on many coasts is probably connected with this fact.

In the summer of 1887, the sea breeze took possession of a narrow strip of the eastern coast of Massachusetts on thirty days, and, judging by the weather of that season, this number may fall somewhat below the normal for the same period in other years. The occurrence of the breeze depends on the general weather of the region; it appears most distinctly on warm, clear, quiet days; and is absent on cool, cloudy and rainy days, and on days with strong winds of any direction. It produces a distinct and agreeable depression of temperature on the coast, but this effect soon weakens and disappears inland. Although the thermometer records fail to indicate the full measure of the breeze's inland advance, observers away from the coast are frequently able to distinguish the sea breeze by its freshness and its faint scent of the sea, which give it a character quite apart from that of the land winds.

The records kept at Manchester, Massachusetts, a station which is near the water's edge and very accessible to the unwarmed sea breeze, show a mean diurnal range on sea-breeze days of 6·9°, while in central Massachusetts the same is 14·2°. A thermograph record from Manchester gives clear indications of the effect of the sea breeze in the truncation of the diurnal temperature curve, two maxima being often apparent, one before and one after the sea breeze. The same feature appears, although with less distinctness, in the thermograph records at Cambridge, Massachusetts, 6 or 8 km. from the shore.

The sea breeze on the coast of Senegambia, at Joal, during the dry winter season, has been described by Bigourdan, who carried on an interesting series of observations there.[1]

[1] *Comptes Rendus*, CXVIII., 1894, 1201-1204. "Résumé des Observations météorologiques faites à Joal (Sénégal) par la Mission chargée par le Bureau des Longitudes d'observer l'Eclipse totale de Soleil du 16 Avril, 1893."

During the stay of the solar eclipse expedition at Joal, from January 1 to April 16, 1893, there were only two or three occasions on which a trace of rain fell. The prevailing wind was northeast. While this wind blows, the temperature and dryness rapidly increase. After noon, the sea breeze sets in, which comes from the northwest, lowers the temperature and increases the humidity. The natives await the coming of this breeze with impatience, but it does not penetrate far inland, and advances very slowly. When the sea breeze approaches from the ocean, the set of the waves changes, as does the colour of the water, as far as the eye can reach. The dividing line can be clearly distinguished at a distance of two to three km., and it usually takes half an hour for the breeze to reach the land. This gives a maximum velocity of advance for the sea breeze of six km. an hour, and also explains the deliberate penetration of the sea breeze inland. During the night and morning, the wind is northeast, and brings in fresh air, but as soon as the sun begins to shine, the air is warmed by contact with the hot ground, and the thermometer rises very rapidly. When the sea breeze is delayed in its beginning until toward 2 or 3 P.M., the temperature may be observed to rise to 40°; but the breeze usually begins before noon, and then the maximum temperature does not exceed 28°-30°. The fall in temperature when the sea breeze sets in is extraordinarily rapid; so rapid, indeed, that the self-recording thermometer cannot keep up with it, and lags several degrees behind. The humidity, on the other hand, rises just as rapidly as the temperature falls. The following data, for April 14, 1893, show what effect the sea breeze has upon the temperature and the humidity at Joal.

TEMPERATURE, RELATIVE HUMIDITY AND WIND DIRECTION AT JOAL, SENEGAMBIA, ON APRIL 14, 1893.

Hour,	6 a.m.	7 a.m.	8 a.m.	9 a.m.	10 a.m.	11 a.m.
Temperature (in degrees),	20·8	23·8	27·3	30·6	33·1	36·8
Relative Humidity,	43%	33%	24%	18%	14%	6%
Wind,	ENE.	ENE.	ENE.	NE.	NE.	NE.
Hour,	12 noon.	12.30 p.m.	12.45 p.m.	1 p.m.	2 p.m.	3 p.m.
Temperature (in degrees),	38·3	39·2	28·0	26·1	25·4	24·0
Relative Humidity,	4%	3%	45%	61%	64%	65%
Wind,	NE.	NE.	NW.	NW.	NW.	NW.

It appears that the arrival of the sea breeze, between 12.30 and 12.45 P.M., causes a fall in temperature of 11°, and a rise in the humidity of 42 per cent. Such sudden changes as these are naturally entirely lost sight of in the regular observations made thrice daily, and can be followed only when self-recording instruments are used.[1]

[1] Reproductions of some of the records of humidity, temperature, and pressure obtained by means of self-recording instruments may be found in the *Annales du Bureau des Longitudes*, V., Paris, 1897.

Relative velocity of land and sea breezes. Diurnal variation in wind velocity on land.—The land breeze, as a rule, seems to be much less well developed than the sea breeze, and Ferrel explains the greater velocity of the latter as compared with the former in the following way. The land is warmer than the water, even when the mean daily temperatures are considered. This is true for the tropics all the year around, and for higher latitudes in summer. Hence the nocturnal temperature gradient, which gives rise to the land breeze, is weaker than the diurnal gradient, which causes the sea breeze. Furthermore, the velocity of the lower strata of air moving over the land is much reduced by friction against the inequalities of the surface; while there is much less friction over the smooth ocean surface. Lastly, as a result of the warming of the land surface, there is the diurnal increase in the velocity of all winds over the land, and this increase must also affect the sea breeze.

Since self-recording anemometers came into use, it has become clear that everywhere, in all climates, over the land and also along shore, the velocity of the wind, no matter what its direction may be, increases from morning to afternoon. The velocity reaches its maximum at about the time of maximum air temperature, and then rapidly decreases, reaching a minimum after midnight. The increase in wind velocity from morning (about 8-9 A.M.) to afternoon is most marked on clear days, and in some climates is so great at certain seasons that the wind rises to a gale in the afternoon, while it decreases in velocity again towards evening, and there is a calm at night. Thus Schweinfurth and Nachtigal describe the extraordinary increase in the intensity of the northeast trade wind in the interior of Africa during the daytime, while the nights are calm.[1]

Espy, and more recently Köppen, whose work was quite independent of that of Espy, have rightly explained the diurnal increase in wind velocity on land as the result of the interchange of air between the upper and the lower strata of the atmosphere during the day, occasioned by the heating of the ground. In consequence of this interchange, the upper strata, which always move much more rapidly than the lower ones, are brought down to the earth's surface to replace the warmed air, which rises.

Lake breezes.—Even large lakes may give rise to off-shore and on-shore breezes. Forel has demonstrated the existence of such breezes in the case of the Lake of Geneva. On the northern shore of that lake, between Ouchy and Rolle,

[1] Further discussion of this phenomenon can be found in the author's treatise: "Die tägliche Periode der Geschwindigkeit und der Richtung des Windes. S.W.A., LXXIX., 2, 1879, 11-96" (*Z.f.M.*, XIV., 1879, 333-349).

the land breeze (*morget*) blows as a strong north wind from between 5 and 7 o'clock in the evening to between 7 and 9 in the morning. This breeze always begins on the land and works out toward the lake, as Forel has hundreds of times had occasion to observe at Morges. The first weak puffs of wind suddenly appear along the shore, and thence work out on to the lake. In the autumn, when the land cools, while the lake is still warm, the air flows toward the lake from all sides in calm weather, and the *morget* then blows almost continuously. The same thing happens to a more marked degree in winter, when the land is covered with snow. The *morget* is, however, then no longer a true land breeze. The lake breeze, *le rebat*, blows during the day from 10 A.M. till 4 P.M., but with less velocity than the *morget*. This lake breeze begins over the open surface of the lake, and advances toward the shore. The *morget* and the *rebat* are true land and lake breezes, according to Forel, and are not mountain winds. The real descending mountain wind, known as *joran*, is much stronger than the *morget*, but reaches only the base of the mountains, not advancing on to the lake.[1]

An excellent example of a lake breeze is described by Hazen.[2] During the summer of 1882, observations were made at the Lake Crib, a water-tower station three miles east of Chicago, in Lake Michigan, and at Chicago, with a view to determining the relation of the winds at the two stations. From this it appeared that the mean hourly direction of the wind at Chicago, during July of that year, indicated a change from almost due east at 1 P.M., through southeast to almost south at 10 P.M.

The theory of land and sea breezes is as follows. During the morning, the land warms more rapidly than the water, and the warmed air over the land expands upward; or, to put it in another way, the pressure aloft rises over the land more than it does over the water. Therefore the air aloft over the land begins to flow off toward the sea, and the pressure rises over the sea, while it falls over the land. Hence it follows that a lower current of air is set up, flowing from the sea toward the land. This is the sea breeze. At night the conditions are reversed. The land cools more rapidly than the water. The cooling of the land involves a decrease in the pressure aloft over the land,[3] and with it an inflow aloft of warmer air from the ocean. Consequently the surface pressure over the land must increase, and that over the ocean must decrease. Thus there results a current of air from the land out to sea, and this is the land breeze. In the morning and evening, between these times of changing winds, there is equilibrium of pressure and a calm prevails.

[1] See A. Forel: *Le Lac Leman*, Basel, 1886, Vol. I., 302.

[2] H. A. Hazen: "Report on Wind Velocities at the Lake Crib and at Chicago, *U.S. Signal Service Notes*, No. VI., 1883.

[3] In connection with this effect of the warming and cooling of the air upon the pressure in the upper strata, reference may be made to the table at the beginning of the section on mountain climate.

The diurnal period is too short to make it possible or any considerable differences in pressure to be established between land and water. Indeed, the differences in pressure from which the land and sea breezes result are so slight that it has been impossible, until recently, to be sure of their existence.[1] Nevertheless, even the annual means of the hourly pressure readings, obtained with self-recording barometers on the coasts and in the interior of England, have shown the above-mentioned differences in pressure at the earth's surface, which set the land and sea breezes in motion. From 10 A.M. until 11 P.M., the coast stations have a relatively higher pressure, while during the night the inland stations have the higher pressure. This should be the case if the air aloft really flows from land to sea by day, and back from sea to land at night. The fact that the sea breeze first begins off-shore, and then works its way in toward the land, probably results from the circumstance that a weaker gradient can put the air in motion over the smooth sea surface than over the land, where there is much more friction. The air over the land cannot overcome the greater amount of friction, and hence cannot move down the gentle gradients offered to it, until the increased warming over the land has brought the small difference in pressure which produces the sea breeze nearer to its maximum value. Moreover, the vertical air currents, which develop with the increased warming of the ground, reduce the effect of the friction.

The author cannot accept the explanation given by Seemann of the off-shore beginning of the sea breeze.[2] According to Seemann's theory, the rapid warming and expansion of the air over the land in the morning exerts a pressure laterally out toward the ocean, which cannot be overcome until later. There does not seem to be the slightest reason for assuming such a lateral pressure of the air. Moreover, it is quite unnecessary to assume that there is such a pressure simply because of this one phenomenon, since the off-shore beginning can be explained in another way. If we recall that the most rapid hourly increase of temperature on a summer forenoon is about 1·2°, and that this rate is still less aloft, *i.e.*, only 0·02° a minute, the air must certainly be supposed to have time enough to expand upward in the natural direction of least resistance.

Monsoons.—The two contrasted sets of atmospheric currents which are due to the reversal of the temperature conditions over land and

[1] The chief discussion of this matter is due to Blanford and Chambers. H. F. Blanford: "Luftdruckdifferenzen beim Wechsel der Land- und Seewinde an der Küste von Bengalen," *Z.f.M.*, XII., 1877, 129-131; and F. Chambers: "Diurnal Variations of the Barometric Pressure in the British Isles," *Quart. Journ. Roy. Met. Soc.*, V., 1878-79, 133-136; *Z.f.M.*, XV., 1880, 195-196.

[2] R. H. Seemann: "Ueber Land- und Seewinde und deren Verlauf," *Das Wetter*, I., 1884, 79-83; 101-105; 124-131.

water in summer and winter, attain a much greater development than the land and sea breezes, by reason of the longer period during which these differences of temperature are in operation. Hence the seasonal winds are of very great climatic importance. In the middle and higher latitudes of the northern hemisphere, these seasonal winds may even entirely overcome the general circulation of the atmosphere, at least so far as the lower strata are concerned. The contrast between equator and poles, or between south and north, is then no longer the dominant control, but this is replaced by the contrast between continent and ocean, or between east and west. Corresponding to the diurnal period of change of the land and sea breezes on the coast, there is an annual period of change of continental and oceanic winds over the continents. There is, however, a marked difference in the degree of development. In the latter case, each of these winds lasts for nearly half a year, and attains a much greater development and velocity than in the case of land and sea breezes. The name *monsoons*, or seasonal winds, is given to those winds which change with the opposite seasons. Every continent gives rise to such monsoons, but in practice the term is applied only to the larger and more dominant winds of this class.

Continental winds of summer.—In order to understand fully the causes of the atmospheric circulation between continents and oceans, which is of such fundamental importance in climatology, we need to consider at somewhat greater length than we have yet done the effect of the warming and cooling of the land upon the distribution of pressure. As the causes which are concerned in producing monsoon winds continue in operation for almost half a year, their effect upon the distribution of pressure is much more important than in the case of land and sea breezes, and may readily be shown directly by means of observations.

The warming of the atmospheric strata over an extended land surface, and over entire continents, as compared with the warming over the oceans, proceeds as follows. The surface is warmed under the influence of insolation during the day; the lower strata of air are likewise directly warmed, and ascend in individual currents, while other, cooler currents descend to replace those which have risen. Thus there results a continuous interchange of air between the upper and the lower strata, which can be detected in the "unsteadiness" of the air that may always be noted when we look across a warmed surface toward distant objects. This play of ascending and descending air currents is interrupted at night, when there

is a rapid cooling of the surface of the ground by radiation, while the higher strata of the atmosphere are not correspondingly cooled. The convectional currents are set up again on the following day, and the high temperatures of the lower strata, which rest on the surface, are thus communicated to higher and higher levels.[1] Furthermore, the air strata are warmed by the radiant heat from the earth itself.

As has already been explained, the air is very diathermanous in regard to luminous solar radiation, but less so as regards radiation from a cold body, like the earth's surface. The radiation from the latter is therefore partially absorbed by the lower layers of air, and helps to warm them. Under the influence of this process of warming, which proceeds from below upward, the atmospheric strata resting upon the land expand; the upper strata are raised by the lower; and the pressure therefore increases aloft. Hence the isobaric surfaces, which are horizontal when the distribution of temperature is uniform, so that the pressure is everywhere equal at the same altitude above sea level, are raised over the warmed land, and depressed toward the cooler ocean. Assuming that it were possible for a person in a balloon to remain continually at the same height above sea level, he would find the pressure at this height constantly rising in going from the coast into the interior of a warmed continent, until he reached the centre of the warmed area. In consequence of this fact, the atmospheric strata aloft have a slope from the continents out toward the cooler ocean, and the

[1] This process, which was deductively inferred by the author several years ago ("Die Gesetze der Temperatur-Aenderung in aufsteigenden Luftströmungen und einige der wichtigsten Folgerungen aus denselben," *Z.f.M.*, IX., 1874, 337-346), has now been fully confirmed by the results of observations made during two balloon ascents, undertaken at night, from Munich, during July, 1893. The temperature increased during the early morning hours (1-3 A.M.), up to a height of 300 m. above the plateau, the temperature at the surface being about 12°, and that aloft 18·5°. From that altitude, the temperature decreased again up to about 900 m. The temperature and the humidity at the latter level showed that the air there had risen from the plateau on the preceding day by reason of its high temperature. The cooling by nocturnal radiation from the surface reached up to about 300 m., but the air aloft had retained the same (potential) temperature with which it had risen from the surface. (L. Sohncke and S. Finsterwalder: "Einige Ergebnisse wissenschaftlicher Fahrten des Münchener Vereines für Luftschiffahrt," *M.Z.*, XI., 1894, 361-376). An ascending current which carries aloft, as one mass, all the air that has been warmed below, as was formerly supposed to be the case, does not, and for obvious reasons cannot, exist.

air slides off in that direction. Thus the air pressure decreases over the interior of the continent, because the mass of air which rests upon the surface there is diminishing. On the other hand, the pressure increases over the ocean, because here there is an additional supply of air aloft. Hence there results, at sea level, a movement of the air from the ocean toward the continent, which is in the opposite direction to the movement aloft. The air flows from the region of higher pressure to that of lower pressure, and the lower air must therefore flow from all sides toward the warmed continent. As these air currents are deflected, in consequence of the earth's rotation, to the right of their direction of motion in the northern hemisphere, and to the left in the southern, we may expect the following winds around the margins of the continents in summer (summer monsoons) :—

	West Coast.	North Coast.	East Coast.	South Coast.
Northern Hemisphere, -	NW.	NE.	SE.	SW.
Southern Hemisphere, -	SW.	NW.	NE.	SE.

In accordance with the usual classification of wind systems, a circulation of air such as that which is developed over a well-warmed continent, is known as a *cyclonic* circulation.

Cyclonic and anticyclonic wind systems.—A cyclonic circulation is developed wherever the air pressure is below that over surrounding regions; an anticyclonic circulation is developed where the pressure is high, and decreases in all directions from the maximum. (See Fig. 4.)

The air currents which are produced around areas of low and of high pressure are the result of two forces. These forces are the force of gravity, which constrains the air to flow from the region of higher to that of lower pressure; and the deflective force of the earth's rotation, which acts to the right of the direction of movement in the northern hemisphere, and to the left in the southern hemisphere. Under the influence of these two forces, the air currents form a whirl, in which the air circulates from right to left, or in a counter-clockwise direction, around the area of low pressure; and from left to right, in a direction with the hands of a clock, around the area of high pressure (in the northern hemisphere). The direction of these whirls is reversed in the southern hemisphere. As is shown in Fig. 4, the differences in pressure are, as a whole, greater, and the distances between the lines of equal pressure (isobars) smaller, in the region of the barometric minimum than in that of the maximum. In the former, therefore, the winds are stronger, and storm velocities prevail; while calms, or light winds, prevail in the latter. The difference in pressure between two places, measured

along the line of most rapid decrease of pressure, *i.e.*, at right angles to the isobars, is called the *baric gradient*. This baric gradient is, in general, a measure of the velocity of wind movement.

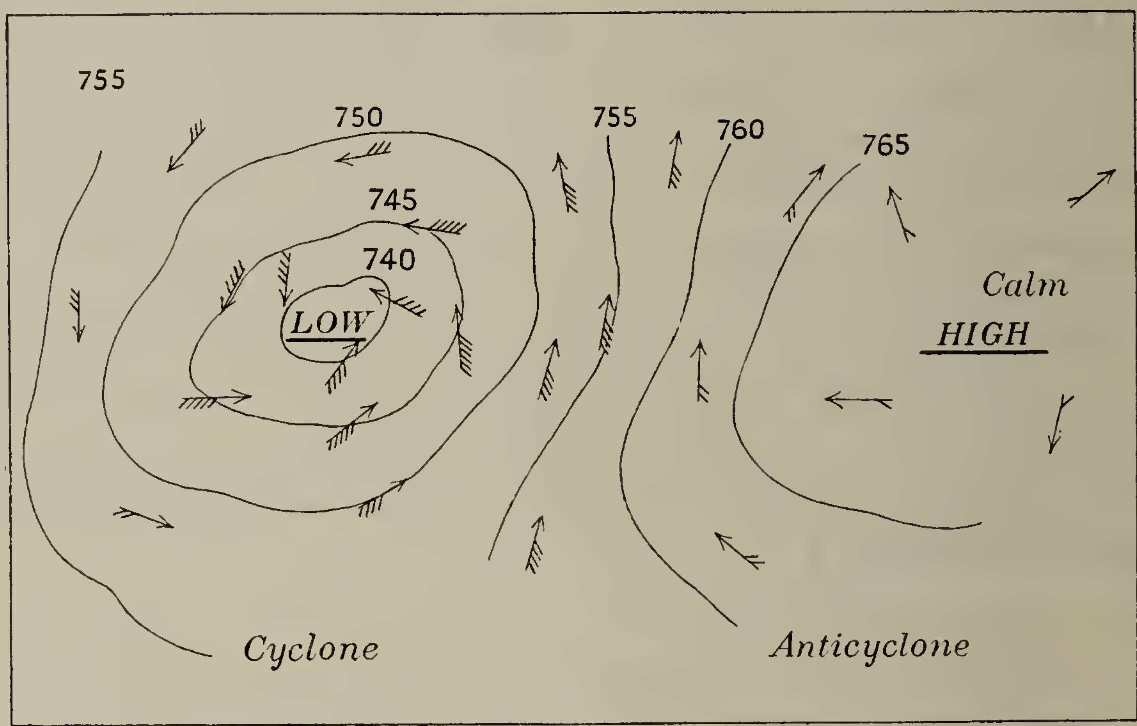

FIG. 4.—CYCLONIC AND ANTICYCLONIC WIND SYSTEMS (NORTHERN HEMISPHERE).

Explanation of the continental winds of summer.—An excellent illustration of the distribution of temperature and pressure, and of the resulting winds, over a warmed land area is given by de Bort for the case of Spain and Portugal in July, when a cyclonic circulation

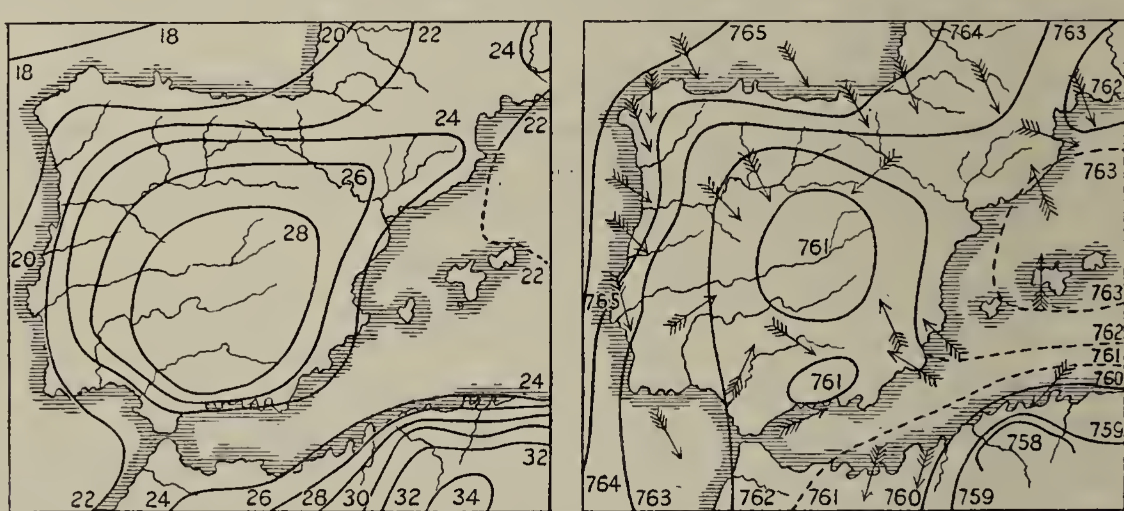

FIG. 5.—ISOTHERMS, ISOBARS, AND WINDS FOR SPAIN AND PORTUGAL IN JULY.

of the winds is developed over the Iberian peninsula and northern Africa.[1] (See Fig. 5.) The largest example of this kind in the world is found over the continent of Eurasia in summer. If the difference of

[1] Teisserenc de Bort: "Etude de la Circulation atmosphérique sur les Continents: Peninsule Iberique," *Ann. Bur. Cent. Met. de France*, 1879, Tome IV., Paris, 1880, 19-60.

temperature between land and water is slight; if the extent of the land area is small, or if there are prevailing winds which result from the general distribution of pressure (as, *e.g.*, the trade winds within the tropics), then the influence of the continents is seen only in a deflection of the prevailing winds, the deflective forces having the directions noted above.

When the air moves in from all sides toward a continent, as is the case over the Iberian peninsula, and as is also seen, on a still larger scale, over the larger continents in summer, the barometric minimum would soon be filled up, and disappear, if the air were not continually moving away in the opposite direction aloft. The reasons for this outward movement aloft have already been considered at some length, and it has also been shown that this is really the primary phenomenon. The movement of the air over a warmed continent is often thought of as a great ascending current, on an enormous scale (*courant ascendant*). This view needs some modification, in so far as an upcast wind, such as is found to exist in mountains, and which is really a *courant ascendant*, does not, and indeed cannot, exist in this case. The air over the land is gradually expanded by the heat throughout its mass, and the upper strata are raised above the level of equilibrium. Hence these upper strata move away aloft on all sides. Now, as the warming continues, new masses of air from below are raised by the process of expansion; they replace the air which has moved away, and thus the movement continues. It probably takes a good deal of time for the individual particles of air, which were on the surface at the start, to reach the higher strata, in which the outward movement begins. These particles do not rise directly from the surface to this altitude, but reach it only by means of a gradual ascent, during which they become involved in ascending and descending air currents. That an immediate ascent to the maximum altitude does not take place is evident from the slow rate of decrease of temperature upwards in clear summer weather; and from the fact that the sky is not continually cloudy in the afternoon. The air, even in the desert, is not so dry that it could rise 3000-4000 m. without producing a cloud cover as the result of its adiabatic cooling. If the air rose all at once, the decrease of temperature would be 1° in 100 m., whereas it is actually only about one-half as large.

Continental winds of winter and their explanation.—The reversed system of air currents is developed over the continents of the middle and the higher latitudes in winter, when the oceans are warmer than the lands. The influence of the land upon the course of the winter

isotherms is very strikingly shown in the distribution of the mean temperatures of the month of December, 1892, over Denmark. There were then reproduced, on a small scale, over Jutland, Fünen, and Zeeland, the same conditions as are presented by the average courses of the isotherms over the great continents of the higher latitudes. Every island and every peninsula had its cold centre, and the temperature everywhere increased toward the sea coast.

Fig. 6.—Isotherms over Denmark, December, 1892.

The difference in temperature between the interior of Jutland and the outlying islands in the North Sea reaches 5° – 6°, in the monthly mean; and between the interior and the islands in the Cattegat, the difference is more than 3°. The temperature minima in the interior of the main land mass reached – 12° to – 16°, while they did not go below – 4·8° over the islands in the North Sea. This, it is to be noted, is not by any means an exceptional case, but is the rule every autumn and winter. Similarly, the isotherms of the British Isles, in December,

1879, and January, 1881, furnish remarkably good illustrations of the cooling of the lands in winter as contrasted with the surrounding water areas.[1]

The course of the isotherms over Denmark, in May, 1892, presents an instructive contrast to the conditions in December. In the former month, the ocean is still cool, but the land warms rapidly. Hence the interior of the land areas is now the warmer, and the temperature

FIG. 7.—ISOTHERMS OVER DENMARK, MAY, 1892.

decreases toward the coast. In the interior of Jutland, the mean temperature is now 10° – 11°, and on the islands it is only 8° – 9°. The North Sea is cooler than the Cattegat, which is well enclosed by land; and here even the islands have a mean temperature of 9°, while the temperature in the interior of Seeland is as high as 11°.

[1] W. Marriott, "On the Frost of December, 1879, over the British Isles," *Quart. Journ. Roy. Met. Soc.*, VI., 1879-80, 102-112, pl. VI., and "The Frost of January, 1881, over the British Isles," *ibid.*, VII., 1880-81, 138-152, pl. VIII.

In winter, when the land is colder than the ocean, the air from the oceans flows in aloft toward the continents. Thus there comes to be an area of low pressure over the water, and a cyclonic movement of the surrounding atmosphere results. Over the continents, on the other hand, the pressure rises because of the inflow of air aloft; an area of high pressure results, and the air consequently flows from land to sea below, ultimately taking part in the cyclonic circulation over the oceans. This process needs further discussion, but in our consideration of it we must confine ourselves to the northern hemisphere, because there are no continents in the higher latitudes of the southern hemisphere.

The flow of air away from a barometric maximum in the northern hemisphere, and the deflection of these air currents to the right, in consequence of the earth's rotation, result in producing northwest winds on the eastern side of a continent; northeast winds on the southern side; southeast winds on the western side; and southwest winds on the northern side. The direction of the circulation of the air around the area of high pressure, is, therefore, from left to right; *i.e.*, with the hands of a clock, and contrary to the direction of movement around an area of low pressure. For this reason, the system of winds around a barometric maximum has been given the name of *anticyclone*.

The maximum pressure over Eurasia in winter is in the neighbourhood of Lake Baikal. The eastern coast of Asia has prevailing northwest winds, which blow towards the barometric minimum over the northern Pacific Ocean, and therefore belong to the left side of a cyclone. Southern Asia has north and northeast winds; Turkestan and southern Siberia have east winds; northern Asia has southwest winds. Europe, on the other hand, with the exception of its southern portion, belongs to the district of the great cyclonic circulation over the North Atlantic Ocean, and therefore has west and southwest winds. The normal development of the anticyclonic circulation on the continent of Eurasia is disturbed by the unsymmetrical distribution of pressure, especially by the tongue of the Atlantic minimum, which extends far into the European polar sea.

North America, like Asia, has, on its eastern coast, prevailing northwest winds in winter, which belong to the rear of the Atlantic barometric minimum. The systematic development of the atmospheric currents on the west coast is prevented by the massive mountain ranges along the coast.

The great increase in the difference of pressure between continents and oceans in middle and higher latitudes from summer to winter,

as shown by comparison of the isobars for July and January, must also be taken into account. In summer, the difference of pressure between the barometric minimum of the interior of Asia and the barometric maximum of the Atlantic and the Pacific Oceans, is about 15 mm. In winter, on the other hand, this difference, with a pressure decrease from land to sea instead of from sea to land, amounts to at least 30 mm. toward the North Atlantic Ocean, and about 25 mm. toward the North Pacific. The difference in winter is, therefore, on the average about twice as great as in summer. This fact can easily be explained in the light of what has already been said concerning the influence of the land upon the temperature of the air. In summer, the temperature increases at a much slower rate from ocean to continent than that at which it decreases in this direction in winter. Indeed, it has already been seen that the difference of temperature at latitude 53° north, between the west coast of Europe and western Siberia is 6·7° in summer and 23·7° in winter. The continent cools much more in winter than it warms in summer. Hence the wind movement between ocean and continent is much more energetic and more extended in winter than in summer. In other words, the slope of the isobaric surfaces in the upper strata of the atmosphere from the ocean toward a large continent is much steeper in winter than is the reversed slope, from the continent to the ocean, in summer. There results from this fact a considerable accumulation of air over the continents in winter, and the production of well-marked areas of high pressure over them; and similarly the production of deep barometric depressions over those portions of the northern oceans which are relatively the warmest. The much more stormy character of the winds, and the greater ranges of temperature in winter than in summer, result from these pressure conditions. The distribution of temperature, and therefore also the distribution of pressure over the hemisphere as a whole, are much more uniform in summer than in winter.

The climatic contrasts of the eastern and western coasts of continents in middle and higher latitudes are explained by what has already been said concerning the prevailing winds of these regions. On the eastern coasts, dry, cold, off-shore winds prevail during the winter. Since these winds come from the colder parts of the continent, the continental climate extends to the coast at this season. It even extends still farther, to the islands lying off-shore, as in the case of the islands of the Japan group, off eastern Asia. In summer, on the other hand, cool, southeasterly, on-shore winds prevail. The coastal districts north of about latitude 45° now have a damp, cloudy marine

climate. Thus these districts are cooled both in winter and summer, and the mean temperatures must therefore be much lower than those of the west coasts, where there is less tendency toward on-shore winds.

The following table illustrates more accurately the relations which have just been described. The first sets of figures in each group give the frequency of the winds in percentages; and the second sets give the effect of these winds upon the mean temperatures, shown in the departures from these means.[1]

FREQUENCY OF WINDS OF DIFFERENT DIRECTIONS, AND THE EFFECT OF THESE WINDS UPON TEMPERATURE.

Western Europe, Eastern Asia, and Eastern North America.

I. Winter—(Frequency).

	N.	N.E.	E.	S.E.	S.	S.W.	W.	N.W.
Western Europe, - -	6	8	9	11	13	25	17	11
Eastern Asia, - -	17	8	5	6	6	8	18	32
Eastern North America,	11	15	6	6	7	18	14	23

Temperature Departure.

	N.	N.E.	E.	S.E.	S.	S.W.	W.	N.W.
Western Europe, - -	– 3·0°	– 3·9°*	– 3·2°	– 1·3°	1·3°	3·1°	2·4°	– 0·4°
Eastern Asia, - -	– 0·6°	0·3°	1·3°	2·8°	3·5°	2·1°	– 0·3°	– 1·2°*
Eastern North America,	– 2·4°	0·6°	3·6°	5·3°	5·8°	4·2°	0·6°	– 2·5°*

II. Summer—(Frequency).

	N.	N.E.	E.	S.E.	S.	S.W.	W.	N.W.
Western Europe, - -	9	8	7	7	10	22	20	17
Eastern Asia, - -	10	9	12	26	16	10	7	10
Eastern North America,	8	12	6	11	13	28	9	13

Temperature Departure.

	N.	N.E.	E.	S.E.	S.	S.W.	W.	N.W.
Western Europe, - -	– 0·1°	0·9°	1·7°	2·2°	1·7°	0·2°	– 1·0°	– 1·0°*
Eastern North America,	– 1·8°	– 1·9°*	– 1·6°	– 0·4°	1·0°	1·2°	0·1°	– 1·2°

[1] For information concerning the method employed in obtaining these means, the reader is referred to the author's treatise, "Untersuchungen über die Winde der nördlichen Hemisphäre, I., II.," *S. W. A.*, LX. 2, 1869, 163-228; and LXIV. 2, 1871, 377-429. The winds of eastern Asia and of North America, here given, have been derived from a more recent computation.

This tabulation shows that in Europe the warmest wind (S.W.) prevails in winter, while in eastern Asia and the eastern United States, the coldest wind (N.W.) prevails. In eastern Asia, in particular, continental winds from north, northwest and west prevail during the winter months, with a constancy almost like that of the trade winds. In summer, the direction of the wind, in western Europe, changes from southwest toward west and northwest. This is not a very marked change, but in eastern Asia, the prevailing wind completely reverses its direction, changing from northwest to southeast. The change in the eastern coastal districts of the United States is similar, but less marked. The accompanying figures, which show these changes in wind direction, were obtained by subtracting the frequency in winter from that in summer. Corresponding data for Russia and western Siberia have been added for purposes of comparison.

CHANGES IN WIND DIRECTION FROM SUMMER TO WINTER ON THE COASTS OF EURASIA AND OF NORTH AMERICA.

Change in Frequency: Summer—Winter.

	N.	N.E.	E.	S.E.	S.	S.W.	W.	N.W.	Sums.
I.									
Western Europe, - -	3	0	−2	−4	−3	−3	3	6	24
Russia, - - - - -	5	1	−1	−6	−4	−3	2	6	28
Western Siberia, - -	6	4	−3	−4	−4	−6	2	6	35
II.									
Eastern Asia, - - - -	−7	1	7	20	10	2	−11	−22	80
Eastern North America, -	−3	−3	0	5	6	10	−5	−10	42

In summer, the wind turns toward the ocean, *i.e.*, to northwest on the western coasts, and southeast on the eastern coasts of the continents. The effect of these winds is everywhere a cooling one, because the off-shore winds are the warmest in summer.

Wind roses for eastern and western coasts.—Figure 8 illustrates the difference in the prevailing winds on the eastern and the western coasts, as well as the effects upon temperature. The length of the radii from the centre of the circles is proportional to the frequency of the winds; the shaded area corresponds to those portions of the wind roses in which the cool winds are found. The size of these shaded areas is therefore proportional to the cooling effect of the winds; and it is clear that this effect is much greater in winter on the

east coasts than on the west coasts. The cooling winds come off the land in winter, *i.e.*, they are northeast on the west coast, and northwest on the east coast. In summer, the cooling winds come from the ocean. The side toward the ocean is also everywhere the rainy side of the wind rose.

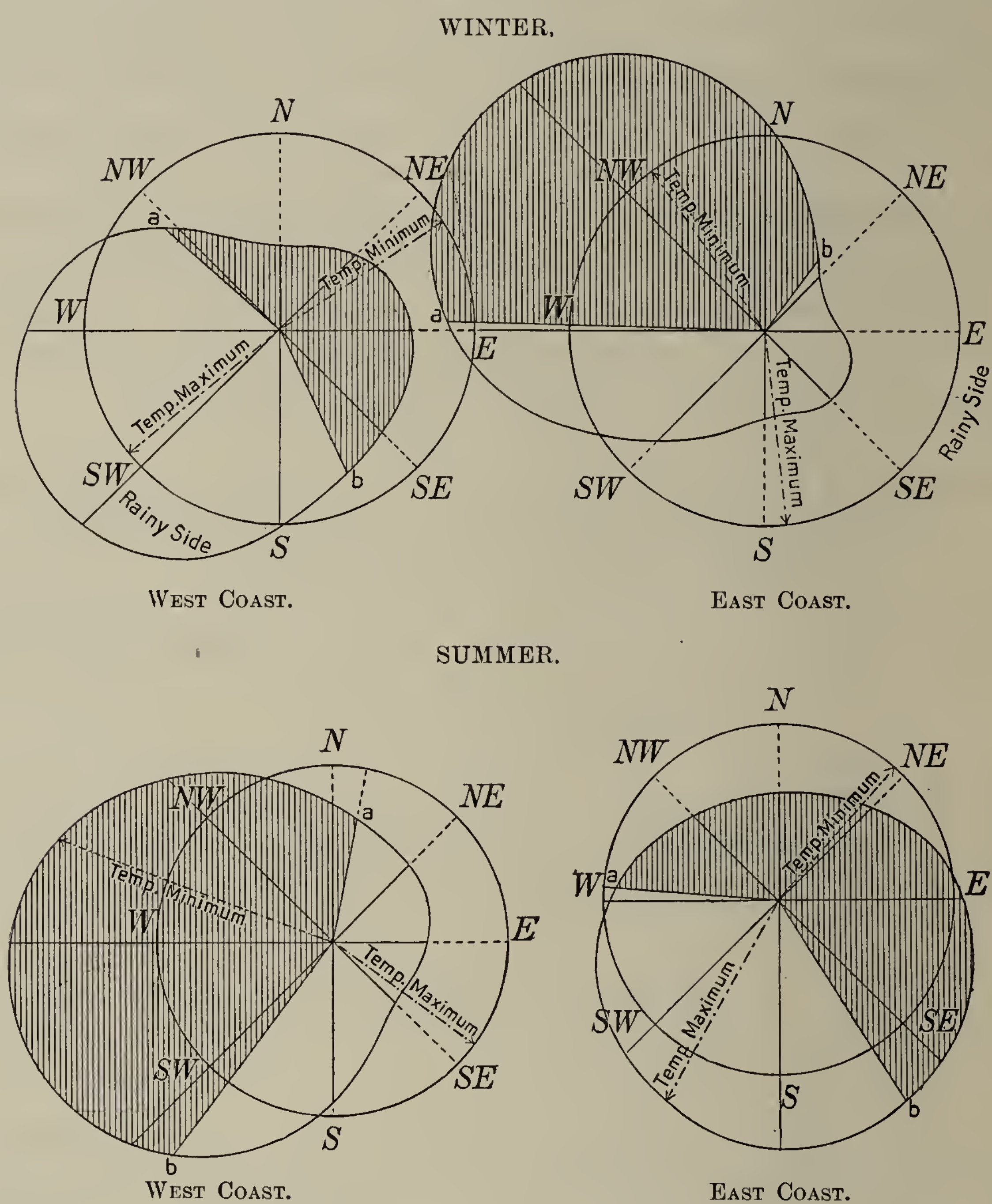

Fig. 8.—Temperature Wind Roses for Eastern and Western Coasts of Eurasia and of North America.

The accompanying table illustrates very clearly the changes in the direction of the coldest and the warmest wind, which, be it said, are in general indicated by the course of the isotherms, as well as the differences of temperature between these winds, noted in passing from the western to the eastern side of the great Eurasian continent.[1]

[1] Woeikof has pointed out that the direction of the coldest wind in the wind roses constructed by the author is not perpendicular to the direction of the isotherms, but lies to the left of the normal to the isotherms. The angle is 22°–45°.

DIRECTION OF COLDEST AND WARMEST WINDS ON THE COASTS OF EURASIA AND OF NORTH AMERICA, AND THE EFFECT OF THESE WINDS ON TEMPERATURE.

I. WINTER.

	N.W. Europe.	Germany.	Central Russia.	Western Siberia.	Eastern Asia.	Eastern North America.
Direction of Coldest Wind, - - -	N. 62° E.	N. 46° E.	N. 26° E.	N.	W. 44° N.	W. 65° N.
Direction of Warmest Wind, - - -	S. 44° W.	S. 55° W.	S. 21° W.	S. 15° W.	E. 84° S.	E. 81° S.
Difference of Temperature, - -	5·6°	7·1°	10·6°	11·1°	4·7°	8·7°

The direction of the coldest wind, in going from the west to the east coast, changes from E.N.E. through N. to N.W.; that of the warmest wind changes only from S.W. to S. The difference of temperature, and hence the changes of temperature with the change of wind, are greatest in the interior of the continent. The high temperature of the S.E. and S. winds in the eastern United States is explained by their coming off the warm Gulf Stream. Since the decrease of temperature per latitude degree between latitude 20° and 60° north is 0·5° on the west coast in January, but is 1·3°, or almost three times as great, on the east coasts, the east coasts must, in general, be exposed to greater temperature changes in winter than the west coasts. The reason that this condition is less marked in eastern Asia is because of the steadiness of the winds there.

DIRECTION OF COLDEST AND WARMEST WINDS ON THE COASTS OF EURASIA AND OF NORTH AMERICA, AND THE EFFECT OF THESE WINDS ON TEMPERATURE.

II. SUMMER.

	N.W. Europe.	Germany.	Central Russia.	Western Siberia.	Eastern North America.
Direction of Coldest Wind, - - -	W. 20° N.	W. 22° N.	W. 53° N.	W. 77° N.	N. 43° E.
Direction of Warmest Wind, - - -	E. 32° S.	E. 45° S.	E. 39° S.	E. 76° S.	S. 29° W.
Difference of Temperature, - -	3·7°	3·4°	3·4°	4·5°	3·3°

The direction of the coldest wind, in going from the west coast to the east coast, changes from W.N.W. through N. to N.E.; that of the warmest wind becomes more southerly in the interior, and changes to S.S.W. on the east coast. The difference in the temperatures of the winds is less in summer, and therefore their influence upon the changes of temperature is also less.

George Forster, in the year 1794, seems to have been the first to recognise the systematic difference of temperature between the east and west coasts. He opposed the idea, which was very prevalent in his day, that North America as a whole is colder than the continent of Eurasia, and called attention to the mild climate of the west coast of North America as contrasted with the climate of eastern Asia. Humboldt later clearly demonstrated the difference of temperature between the east coast of North America and the west coast of Europe.[1] The comparison which was made by Humboldt is here repeated, with some additional data, and with the use of new and more accurate mean temperatures than were available at the time when the original table was compiled.

TEMPERATURES ON THE EASTERN AND WESTERN COASTS OF THE NORTH ATLANTIC OCEAN.

Station.	Latitude.	Mean Annual Temperature.	Coldest Month.	Warmest Month.	Difference in Mean Annual Temperature.
Nain (Labrador), - -	57·2°	– 3·8°	– 19·9°	10·6°	
Aberdeen (Scotland), -	57·2°	8·2°	2·9°	14·3°	12·0°
St. Johns (Newfoundland),	47·6°	4·5°	– 5·3°	15·3°	
Brest (France), - -	48·4°	12·0°	6·6°	18·2°	7·5°
Halifax (Nova Scotia), -	44·7°	6·3°	– 5·2°	18·0°	
Bordeaux (France), - -	44·8°	12·8°	5·8°	20·6°	6·5°
New York (N.Y.), - -	40·8°	10·6°	– 1·7°	24·2°	
Naples (Italy), - -	40·8°	16·5°	9·0°	25·1°	5·9°
Norfolk (Virginia), - -	36·8°	15·1°	4·6°	25·9°	
San Fernando (Spain), -	36·5°	17·5°	11·5°	24·5°	2·4°

The difference in temperature between the opposite shores of the Atlantic Ocean decreases with the latitude, and at about latitude 30° north, the southern United States have nearly the same mean

[1] *Zentralasien*, Vol. II.

temperatures as northern Africa. It will be noticed that, at least as far as latitude 40°, the summer temperatures of the American coast are also lower than those of the same latitude in Europe. It is therefore evident that, in the higher latitudes, the coast of North America has a continental climate in winter, and a modified marine climate in summer. Hence there is a negative temperature anomaly in both seasons, which explains the low mean annual temperature.

A similar contrast of mean annual temperatures is also found between the two opposite shores of the North Pacific Ocean. Under the influence of the almost constant off-shore northwest winds, the winter temperatures of the eastern coast of Asia are reduced still more than are those of North America. The summer temperatures of the Asiatic coast, on the other hand, are higher, while on the northwestern coast of North America they are much lower, than those of the European coast. The following data illustrate these temperature conditions more clearly.

TEMPERATURES ON THE EASTERN AND WESTERN COASTS OF THE NORTH PACIFIC OCEAN.

Station.	Latitude.	Mean Annual Temperature.	Coldest Month.	Warmest Month.	Difference between Mean Annual Temperatures.
Ayan (Siberia), - -	56·5°	–3·9°	–20·4°	12·4°	
Sitka (Alaska), - -	57·1°	5·7°	–1·0°	12·6°	9·6
Nikolaievsk (Siberia), -	53·2°	–2·5°	–22·9°	16·4°	
Fort Tongas (Alaska), -	54·8°	8·1°	1·1°	15·1°	10·6
Vladivostok (Siberia), -	43·2°	4·6°	–15·0°	20·8°	
Fort Umpqua (Oregon), -	43·7°	11·4°	6·8°	15·5°	6·8
Peking (China), - -	39·9°	11·8°	–4·6°	26·2°	
Marysville (California), -	39·2°	16·4°	7·4°	25·3°	4·6
Shanghai (China), - -	31·2°	15·7°	3·2°	28·2°	
San Diego (California), -	32·7°	16·7°	11·9°	22·2°	1·0

The difference in the winter temperatures is here seen to be more than 20° between latitudes 40° and 60° north, and almost 9° at latitude 32°. The eastern coasts of the great continents of the northern hemisphere resemble one another in having an extreme range of temperature from winter to summer. These coasts, especially the eastern coast of Asia, are therefore to be classed among the severe

climates; while the western coasts have a limited marine climate, with small annual ranges of temperature.

During the prevalence of the off-shore winds of winter, the eastern coasts also have certain other important characteristics of a continental climate, namely a low relative humidity; little cloudiness, and a deficiency of rainfall. The latter, however, is noted only on the eastern coast of Asia. The prevailing summer rains, which, in eastern Asia, constitute an actual rainy season as far as the higher latitudes, are a characteristic of eastern coasts and of continental interiors in general. In consequence of the on-shore winds which prevail throughout the year, the western coasts have high relative humidity and much cloudiness, and their rainfall is more evenly distributed throughout the year.

Influence of bodies of water upon temperature and humidity.—The climatic contrasts which have been seen to exist between the eastern and the western coasts of the continents north of about latitude 30° N. are repeated, on a smaller scale, on the eastern and western coasts of peninsulas and of large islands lying near continents, and also on the eastern and western coasts of inland seas in middle and higher latitudes. In winter, there is everywhere a tendency to the formation of a barometric minimum over the enclosed portions of the oceans, and the occurrence of southerly and westerly winds on the eastern sides, and of northerly off-shore winds on the western sides of large bodies of water. These conditions reproduce, on a small scale, the larger effects upon climate which have already been considered. This influence is especially effective when some peculiar distribution of pressure prevents the prevailing movements of the atmosphere from controlling that particular portion of the earth's surface. Under these circumstances, the local conditions may develop a system of wind movements of their own, as Hoffmeyer has shown by several instructive examples.[1]

Water areas have less effect in summer, and their effect is then of the opposite character. The proximity of an extended water surface tends to lower the temperatures at that season. In spring, also, such a surface has a similar cooling effect (see Fig. 8). This is particularly true when the water is frozen over, as happens in the winter of the higher latitudes, for the warming in spring is then considerably delayed. In autumn, on the other hand, because of their slower cooling, water areas tend to keep the temperatures higher than they otherwise would be. The proximity of cold water surfaces has a notable influence upon atmospheric humidity in the spring months, for it tends

[1] N. Hoffmeyer: "Weitere Bemerkungen über die Luftdruckvertheilung im Winter," *Z.f.M.*, XIV., 1879, 73-82.

to keep the air dry, and thus favours the occurrence of late frosts. In autumn the reverse is true.

An interesting illustration of the warming of cold air currents in winter as a result of their blowing over an open water surface may here be noted. Attention was called to this case by Köppen, who observed it on the European daily weather maps of January, 1894. After a fall of snow, a spell of severe cold weather followed, which extended over the interior of England and of Ireland. The cold east winds which blew off the continent of Europe during this period, had a considerably higher temperature on the coast of England, after they had crossed the North Sea. (Helder, on the coast of Holland, is opposite Yarmouth, on the eastern coast of England.)

WIND DIRECTION AND VELOCITY, AND TEMPERATURE AT HELDER AND YARMOUTH ON JAN. 3-7, 1894.

	Jan. 3.	Jan. 4.	Jan. 5.	Jan. 6.	Jan. 7.
Helder, - -	E. 5, – 3°	E. 8, – 7°	E. 6, – 13°	E. 2, – 8°	S.S.W. 1, –13°
Yarmouth, - -	E. 6, – 1°	E. 9, – 3°	E.S.E. 7, – 6°	E. 5, 0°	E.S.E. 4, 0°

The mean temperature of the east winds at Helder averaged – 7°; at Yarmouth, after they had crossed the North Sea, it was – 2°. These winds were even more warmed farther north, at latitude 55°, where the North Sea has its greatest breadth. At Shields, for example, the temperature was + 1·3°. On January 6 and 7, the wind blew out from the interior of the islands, and the temperature was then – 10°, with a southwest wind. The east winds became cold again, on the west coast of England, at Liverpool; for the mean temperature for January 3-7 was – 4° at that station. On January 6, the E.S.E. wind, at Liverpool, had a temperature of – 9°, while its temperature on the eastern coast was only 0°. In the interior of Ireland, at Parsonstown, the temperature fell to – 14° during a calm and under a clear sky. The winds blowing from the interior brought frost to the west coast, and even Valentia had a temperature of – 6° on January 6. These temperatures were all taken at 8 a.m.[1]

Lake Ontario tempers the cold waves of winter in a similar way, as may be seen by the following example. On January 19, 1892, an anticyclonic area passed eastward across Canada, giving northerly winds

[1] W. Köppen: "Die Kälte der ersten Januarwoche auf den britischen Inseln," *Das Wetter*, XI., 1894, 29-31.

and very cold weather over the northeastern States. The temperatures observed on the northern and southern shores of the lake are given in the accompanying table.

EFFECT OF LAKE ONTARIO IN TEMPERING COLD WAVES IN WINTER.

Date.	North Shore.		South Shore.	
	Kingston.	Toronto.	Oswego.	Rochester.
Jan. 19, 8 a.m., - -	−16·7°	−15·6°	−11·1°	−8·9°
Jan. 19, 8 p.m., - -	−21·1°	−17·8°	−13·3°	−11·1°
Jan. 20, 8 a.m., - -	−30·0°	−22·2°	−18·9°	−14·4°

Kingston is near the head of the St. Lawrence river, and at a greater distance from the lake than Toronto. The winds were northerly throughout the observations. In this case, the lake appears to have maintained the temperature at Rochester from 7·8° to 15·6° above the point to which it would otherwise have fallen. The frequent occurrence of such conditions gives the south shore of Lake Ontario an average midwinter temperature 2·8° higher than that of the north shore.[1]

Every lake, and indeed every large river, has effects upon the temperatures in its vicinity similar to those above considered, although these effects are very slight, and can usually not be demonstrated by means of meteorological observations. The higher degree of atmospheric humidity in the vicinity of bodies of water causes a more abundant deposition of dew in summer, and more frequent fogs, especially in spring and autumn, when the surface temperatures of the water lag behind those on shore; and when there is consequently a frequent intermingling of warmer and cooler air.

[1] E. T. Turner: *The Climate of the State of New York.* Fifth Annual Rept. Met. Bur. and Weather Service of the State of N.Y., 1893. 8vo, Albany, 1894, p. 370.

CHAPTER X.

INFLUENCE OF OCEAN CURRENTS UPON CLIMATE.

Winds and ocean currents.—The difference in temperature between the eastern and the western coasts of continents, which has been discussed in preceding chapters, is not determined solely by the prevailing winds, for the ocean currents also play an important part. This is especially true in the lower latitudes, where the differences in temperature between eastern and western coasts are completely reversed under the influence of these currents. The great ocean currents are caused by the prevailing winds, and for this reason the general direction of flow of the ocean currents agrees with the direction of the prevailing winds over the corresponding portions of the oceans.[1] Hence the ocean currents themselves may be considered meteorological phenomena.

A diagrammatic representation of the surface currents of the oceans on both sides of the equator is given in the following figure (Fig. 9), which is reproduced from Wild's *Thalassa*. If this figure be compared with charts showing the distribution of pressures and winds[2] it will be seen that the atmospheric currents, as well as the ocean currents, circulate around the centres of high or low pressure over the oceans. The barometric minimum of higher latitudes is encircled by currents in atmosphere and ocean, flowing from right to left, the cold polar currents being on the left, and the warm equatorial currents being on the right. The barometric maximum in the vicinity of latitude 30° has a circulation about it from left to right, the cool currents being on the right, and the warm currents on the left. These statements hold only

[1] J. Hann: *Allgemeine Erdkunde*, 5th Edition, 1896, 274.

[2] J. Hann: *Atlas der Meteorologie*; or Bartholomew's *Atlas of Meteorology*, Plates 12 and 14.

for the northern hemisphere. In the southern, these systems of winds and currents move around in the opposite way. The effect, however, is always the same; for the western coasts, in low latitudes, are everywhere cooled, and the eastern coasts are warmed, while this condition is just reversed in higher latitudes.

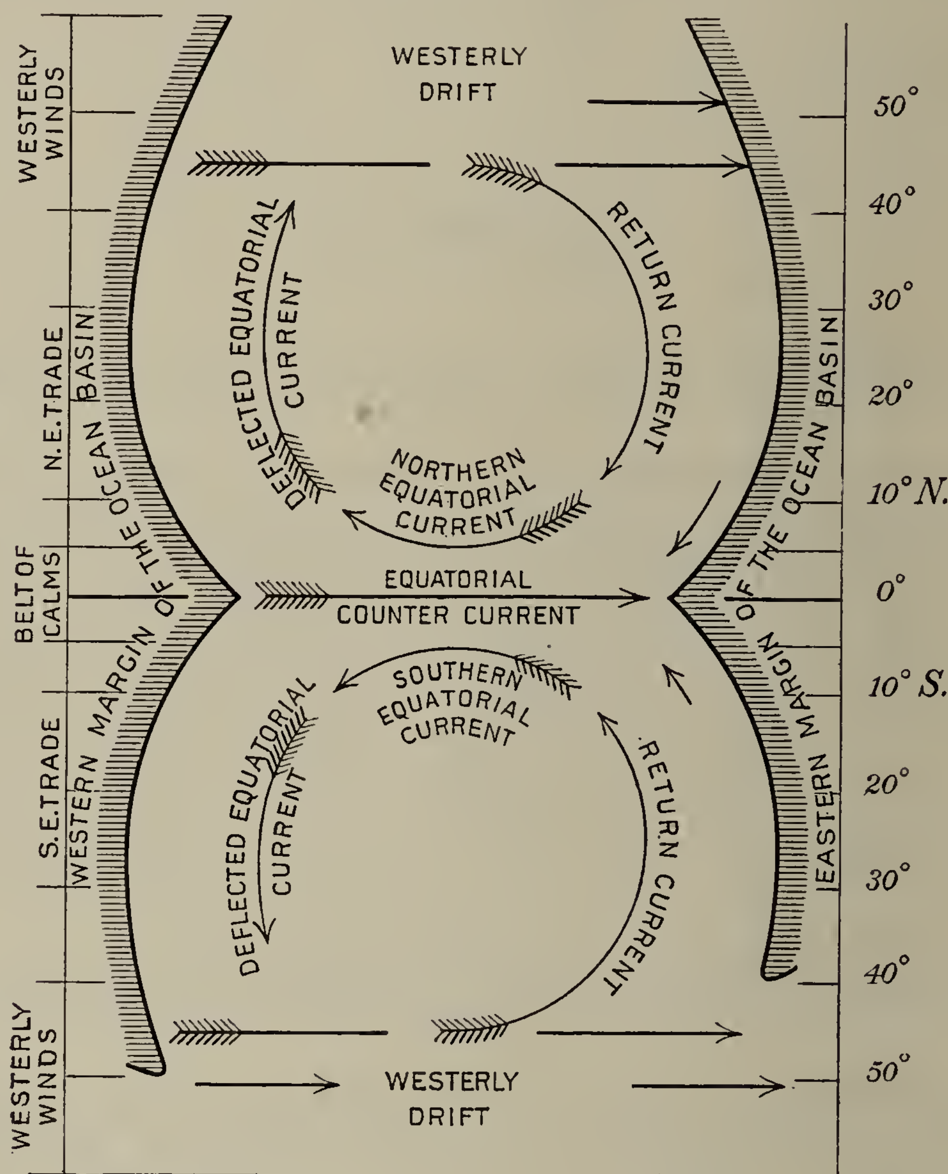

Fig. 9.—Ocean Currents (After Wild).

The ocean currents, as a whole, are thus seen to follow the prevailing winds. The warm currents off the eastern coasts of the continents, namely, the Gulf Stream and the Japan Current, or Kuro Siwo, are an exception, in so far as the cold northwest winds of winter blow offshore, almost directly across these currents, without turning them to any considerable extent from their course. In connection with this anomaly, it need only be noted briefly at this point, that these warm equatorial currents off the eastern coasts are not by any means wind-drift currents. They are, in fact, true ocean rivers, which have their source, and gain their momentum, in the equatorial current produced by the trade winds.

It must be remembered that the equatorial current occupies about twenty degrees of latitude, on both sides of the equator, which is one-third of the earth's surface, and that it embraces certainly more than one-half of the ocean surface in the northern hemisphere. With this in mind, there is nothing surprising in the fact that this great volume of water, when once set in motion, should not lose its momentum as soon as it meets the obstruction caused by the eastern coasts of the continents; but that, turning north and south toward higher latitudes, it should, for some distance, retain enough velocity to keep on its course, even when this course does not everywhere coincide with the direction of the surface winds. When, however, these branches of the equatorial current have poured their warm waters into the farther portions of the ocean basins, in about latitude 40° N. and S., this warm water comes under the influence of the westerly winds which prevail there, and is driven to still higher latitudes. As the result of this movement, the warm water is pushed toward the western coasts of the continents, and the only place that is left for the cold surface currents returning from the polar oceans is on the eastern coasts. In the southern oceans, where there are no great continental masses beyond latitude 40° S., the cold Antarctic currents, which often carry icebergs, have plenty of room.

The ocean currents between the equator and latitude 40° north and south are therefore the following: On the side towards the equator there is the equatorial current; on the eastern coasts of the continents we have the warm branch of the deflected equatorial current, which continues to higher latitudes and there, at about latitude 40°, turns to the right in the northern, and to the left in the southern hemisphere, flowing in an easterly direction poleward from the horse latitudes, and gradually losing its velocity and its high temperature. At this point, the water comes under the influence of the winds which prevail on the eastern side of the tropical high pressure area, and it is again driven equatorward toward the southeast and the south.[1] The current thus makes a turn and again becomes a part of the equatorial current, completing the shorter circuit. This return current on the sub-tropical and tropical western coasts of the continents is cool, for we here have water which has already been cooled, flowing into lower and warmer latitudes.

The effect of these currents upon temperature.—This short tropical and sub-tropical circuit of ocean currents which is developed in the

[1] In the northern hemisphere. In the southern hemisphere this direction is towards NE. and N.

Atlantic and Pacific on both sides of the equator, has an effect upon temperature, which may briefly be described as follows. The temperatures on the eastern coasts in tropical and sub-tropical latitudes are higher than on the western coasts, and there is for a time an increasing negative temperature anomaly when approaching the equator on the west coasts, which are under the influence of the cool return currents from higher latitudes. In illustration of what has just been said, it may here be briefly noted that, in the Atlantic Ocean, the western coast of north Africa (Morocco), and, to a still greater degree, the western coast of south Africa, are abnormally cool. The same thing is true of the California coast, and especially of the coasts of northern Chile and Peru in the Pacific Ocean. The configuration of the coast increases, or diminishes, the amount of the cooling. When the coast bulges out toward the equator, as happens in the case of south Africa and of South America, the cold current hugs the coast. The opposite condition, and the opposite effect, are seen on the coast of North and of Central America. The equatorial region proper, which is usually included in the belt of calms, alone has no share in this cooling. Furthermore, the returning equatorial current is found here, and this is a warm current.

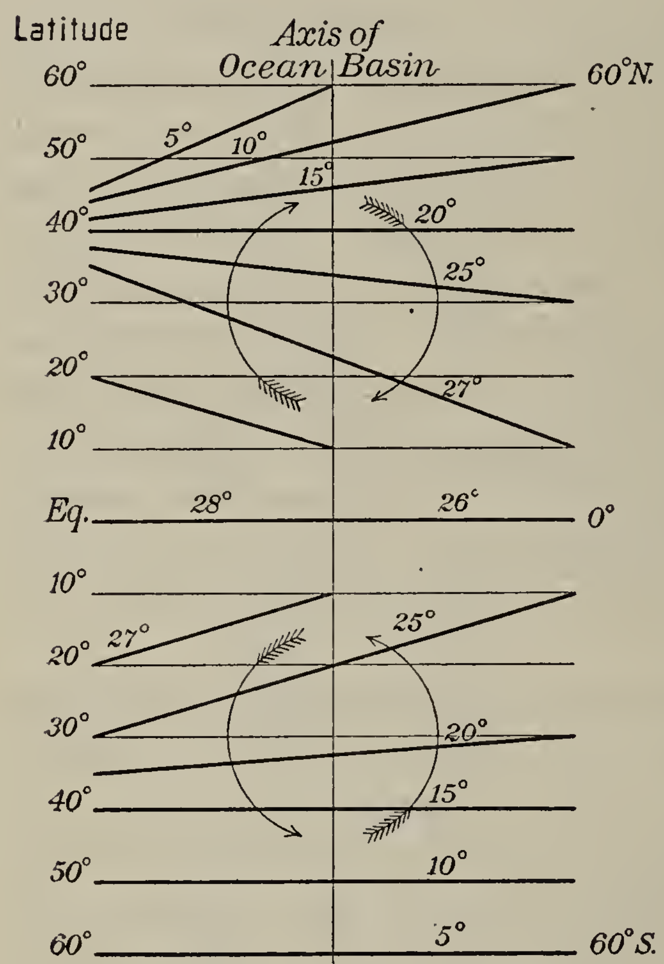

Fig. 10.—Ocean Surface Isotherms (After Wild).

The eastern coasts are relatively warm, as for example, the northern coast of Brazil and the Guiana coast, and the West Indies and the archipelago to the east of Asia. The temperatures in the eastern portions of the tropical oceans are low, because the cool currents here join with the equatorial current. The temperatures are higher in the western portions, because the water has already become warmed during its passage across the ocean, under the heat of the tropical sun. The isotherms of the ocean surface are diagrammatically represented in Fig. 10, which is reproduced from Wild's *Thalassa* (1877).

The effect of cold ocean water near shore upon climate.—In the explanation of the low temperature off the west coasts of South America and of south Africa, as well as off the northwestern coast of northern Africa, one point was formerly overlooked which has been brought to light by recent oceanographic studies, and which makes this phenomenon much more intelligible.[1]

There is one difficulty in the way of referring the cold water off the coast of northern Chile and of Peru, and the cooling of the air which is associated with it, wholly to the ocean current from the south. This difficulty arises from the notable fact that the water temperature at Callao (lat. 12° S.) is no higher than that at Valparaiso (lat. 33° S.). As the current moves at the rate of only 15 nautical miles a day, and therefore takes about four months to go from Valparaiso to Callao, it certainly would have time to become considerably warmed on the way. Captain Dinklage found the coldest water at Callao close to the shore, where there was no perceptible current. The increase of temperature from the shore westward out to sea was found by Captain Hoffmann to be as follows:—

OCEAN SURFACE TEMPERATURES OFF CALLAO, ON THE COAST OF PERU.

	Coast at Callao.	Pacific Ocean—Nautical Miles from Shore.			
		30	80	110	135
Water Temperature, -	18·2°	20·6°	23·8°	26·2°	27·0°

In the light of these facts, Dinklage came to the conclusion that this low temperature may be due to a rise of deep ocean waters along the coast.[2] While the trade wind out over the ocean is driving the water along with it, and is thus producing the equatorial current, there must be a compensating rise of the water along the coast; and this rising body of water brings up with it the low and very uniform temperature of the ocean depths. Where this suction-effect of the trade wind drift is most marked, there will be found the lowest temperature of the water along the coast. For this reason, it is not surprising to find very low temperatures even close to the equator. Furthermore, a sharp

[1] The subject has been discussed by Krümmel, Hoffmann, and Buchanan.

[2] See also "Das Kalte Küstenwasser; Entdeckung der Ursache desselben," *M.Z.*, XVI., 1899, 313.

deflection of an ocean current off-shore may cause a rise of cold water from below. Murray has carried out a series of interesting observations on the Scotch lakes, which clearly show how the wind, by driving and banking up the water, causes a rise of cold water to leeward, and an accumulation of the warm surface water on the windward shores.[1]

This process must be an important one along the shores of all the oceans within the region of constant trade winds. The trades blow the warm water away from the west coasts, and cold water rises to replace that which has been thus removed. The warm water, in consequence, accumulates along the eastern coasts in the tropics, and hence the isothermal surfaces within the tropical, and to some extent also within the sub-tropical oceans, slope from east to west. Cold water is found along the west coast of northern Africa from the Strait of Gibraltar to Cape Verde; along the west coast of southern Africa, from about latitude 10° S. to the mouth of the Orange River, and occasionally as far as the vicinity of Cape Town; along the whole California coast in western North America, and in South America from about latitude 40° S. on the coast of Chile, as far north as Payta, close to the equator, on the coast of Peru.

A typical occurrence of cold water along shore is found on the Somali coast of the northeastern point of Africa, from Cape Warshekh to Cape Guardafui.[2] The cold water is found here during the blowing of the southwest monsoon, from June to September. This monsoon, it will be noted, is an off-shore wind, and wherever strong prevailing winds blow off-shore there results a rise of cold water along the coast. During the season of the southwest monsoon, a mean ocean temperature of 16° is found, in August, at Cape Ras Hafun, in latitude 10° N.; while north of Cape Guardafui, in latitude 12° N., the ocean temperature is 27° to 30°. This cold water along shore, which comes from below, and can readily be detected by its green colour, has a notable effect upon the climate of the adjacent coast. It greatly reduces the air temperature; frequently produces dense fogs, called *garuas* on the coast of Peru, and *cacimbos* on the coast of Benguela; and also leads to a lack of heavy rainfalls, and to the absence of thunderstorms. Under these conditions, the air over the land is dry, and the nights are cool. For these

[1] J. Murray: "On the Effects of Winds on the Distribution of Temperature in the Sea- and Fresh-Water Lochs of the West of Scotland," *Scot. Geogr. Mag.*, IV., 1888, 345-365.

[2] See map in J. Y. Buchanan's "Similarities in the Physical Geography of the Great Oceans," *Proc. Roy. Geogr. Soc.*, VIII., 1886, 753-768.

reasons, the coolest season on the Somali coast is the period of the southwest monsoon, although this really corresponds to the summer.

Captain Hoffmann says: "At the beginning of July, the temperature of the air and of the ocean remained fairly constant between Zanzibar and Cape Warshekh, the water temperature being 25°. The water temperature fell rapidly between latitudes 4° and 8° N., and reached the abnormally low reading of 15° off Ras al Khyle. At the same time, the temperature of the air fell, and even with a clear sky, the thermometer did not rise above 20° at noon; so that it was pleasant to sit out under the tropical sun. The horizon was hazy, and heavy dew formed at night. The colour of the ocean was a dark olive green, and often almost black, while it was light green close to the shore. The water was always deep blue where it had its ordinary temperature. The temperature was found to be uniformly between 15·5° and 15·3° to depths of 200 meters."[1]

The ocean currents north of latitude 40° must next be considered. The regions covered by the sub-tropical barometric maxima over the oceans, between latitudes 25° and 40°, are marked by weak variable winds. These are called the *horse latitudes* by seamen. On the northern side of the tropical high pressure belts of these horse latitudes (in the northern hemisphere), southwesterly winds prevail over the ocean, especially in winter, when a deep barometric depression forms over the northern portion of the oceans, in about latitude 60° N. These southwesterly winds carry the warm water, which was brought to middle latitudes by the Gulf Stream and the Kuro Siwo, to the northeast, as a great drift current, and the western coasts in the higher latitudes are thus warmed. Both air and ocean currents therefore combine to raise the temperature on these western coasts, especially in winter. In the case of the eastern coasts, on the other hand, the warm branch of the equatorial current, which flows close along the coast up to about latitude 40°, does not benefit the continent in winter because the prevailing winds blow off-shore. Thus it is clear that a high ocean temperature can affect the temperature over the land only when the prevailing winds blow on-shore. This is the case on the eastern coasts in summer, but then the ocean is cooler than the land, and at this season the warm current can help to raise the mean temperature only by preventing a considerable cooling through the on-shore winds. Beyond latitude 40° N., where the warm current turns toward the northeast and east, and leaves the coast, we find cold currents on the east coasts, which considerably reduce the summer temperatures. In the Atlantic Ocean this cold current brings down icebergs.

[1] Hoffmann: "Reise S.M.Kr. 'Möwe' von Zanzibar nach Aden," *Ann. der Hydrogr.*, XIV., 1886, 391-396.

The high temperature of the North Atlantic Ocean.—There are two striking features in connection with the ocean currents north of the 40th parallel of latitude, both of which have important climatic consequences. Hence their controls need some explanation. The first of these features is the high temperature of the North Atlantic. This ocean has, both on its surface and down to a great depth, the highest temperatures to be found in its latitudes; indeed, its bottom temperatures are lower than those found in any other ocean. The second feature is the great cold currents which flow up the west coasts of South America and of southern Africa, and join with the equatorial current.

The cause of both these phenomena is to be found partly in the configuration of the continents, or rather in the outline of the eastern coasts north of the equator, and partly in the higher velocity and greater constancy of the southeast than of the northeast trade. It results from these facts that the great body of warm water in the equatorial current is, even at the very beginning of its circuit, driven over toward the northern hemisphere, and, since the barrier which turns this current aside into higher latitudes trends southeast and northwest, the greater portion of the warm water is deflected into the northern hemisphere. This southeast and northwest trend of the eastern continental coast is particularly distinct in the case of South America from Cape San Roque to Trinidad; but it is also seen, to a less marked degree, in the island groups of New Guinea and the Philippines. The northern branches of this warm equatorial current, notably the Gulf Stream and the Antillean current, which accompanies the Gulf Stream on its eastern side, and also the Kuro Siwo, are therefore larger than the warm currents off the southern coast of Brazil and the eastern coast of Australia. The smaller body of warm water in the Kuro Siwo is more or less lost in the broad basin of the North Pacific. On the other hand, the mighty Gulf Stream, with the Antillean current, pours its warmed waters into the ocean basin of the North Atlantic, which quickly narrows north of 40° N.; here, consequently, there is a heaping of warm water that has no parallel elsewhere. The warmer the northern ocean waters, the deeper becomes the barometric depression which forms over this ocean during the winter, and indeed during the greater portion of the year; and hence the stronger and more constant are the west and the southwest winds which drive the warm water along the west coasts into higher latitudes. For this reason, northwestern Europe enjoys the mildest winter climate, and the highest mean temperatures which are found

anywhere in corresponding latitudes. Several factors have thus been seen to combine to give the western coast of northern Europe its extraordinarily favourable climate.

The cold currents off the western coasts of South America and Africa.—As the southeast trade wind covers a broader area of the oceans, and as it blows more steadily and with a higher velocity than the northeast trade wind, the southern equatorial current is stronger than the northern. The former current, therefore, needs a larger supply of water. The movement of the water equatorward under the influence of the southeast trade may be said to give rise to a kind of suction-effect, which draws the water along the tropical western coasts from higher latitudes to help supply the equatorial current. The water thus drawn into the circulation follows the prevailing winds on the eastern side of the sub-tropical high pressure area, and so it is easy to explain the origin of this circulation. The cold ocean currents found off the west coasts of South America and of southern Africa, and flowing toward lower latitudes, have occasionally been given the name of *Antarctic currents*, which is quite unfitting. They have, to be sure, the same direction of flow as the true polar currents, such as the Labrador current and the current off the eastern coast of Greenland, but their origin is not the same. The true analogy is found off the coasts of California and of northern Africa. The Peruvian current, and that off the west coast of southern Africa, are much better developed than the corresponding currents north of the equator, because the southern equatorial current is stronger than the northern. Furthermore, the westerly drift in the southern hemisphere, in which the former currents have their origin, is very well marked because of the strong prevailing westerly winds, and the outline of the southern coasts of south Africa and of South America is very favourable for deflecting this ocean drift toward lower latitudes.

The low temperatures of the southern oceans in higher latitudes, as compared with those north of the equator, result from the narrowing of the continents of the southern hemisphere to the south, while in the northern hemisphere the continents become wider to the north, and enclose the ocean basins. In the southern hemisphere, the smaller body of warm water is spread far and wide over the vast expanse of the southern oceans. North of the equator, the greater volume of warm water carried poleward by the Gulf Stream and the Kuro Siwo is hemmed in by an ocean basin which narrows towards the north. In addition to this, there comes the further fact that the northern ocean basins are either partially, or almost completely, shut off from the water

of the polar seas, while in the southern hemisphere there is no barrier against the cooling effects of Antarctic water and ice. All of these conditions are of great importance if a general comparison is to be made between the climates of the northern and the southern hemispheres.

The effect of ocean currents upon the distribution of rainfall still remains to be considered in a general way. In this matter the temperature of the currents plays an important part. Warm ocean currents, *i.e.*, those which flow from lower to higher latitudes, must increase precipitation on the neighbouring coasts, for the air over the water is saturated with water vapour at a higher temperature than belongs to the particular latitude in which it happens to be. Cold currents, on the other hand, *i.e.*, those which flow from higher to lower latitudes, diminish precipitation in their neighbourhood, because the moist air over them has a temperature which is below that normal to the latitude. The air in the latter case becomes warmed over the land, and its temperature departs more and more from the dew point. Furthermore, these cold currents are also accompanied by winds which blow from higher to lower latitudes, and hence also have the tendency to decrease, rather than to produce, rainfall.

The facts afford a complete confirmation of these general deductions. The cool return currents of the sub-tropical and the tropical system on the east side of the barometric maxima, give rise to a notably small rainfall along the coasts of the continents past which they flow. This deficiency of rainfall is greatest on the west coasts of south Africa and of South America, along which the greatest of these currents flow. The west coast of South America, from the point at which the westerly drift current turns northward toward lower latitudes, has a rainfall which becomes smaller and smaller until an absolutely rainless district is reached, on the coast of northern Chile and of Peru. This is associated with the fact that the negative temperature anomaly becomes more and more marked the farther the current advances into lower latitudes; while the temperature of the water rises very slowly. The rainless belt continues until the cold current leaves the coast. The part played by the cold water along shore in bringing about a deficiency of rainfall has already been discussed. Even when the prevailing south and southwest winds along shore carry the moist air from the ocean on to the land, rainfall can seldom, or never, occur. The reason for this is that the land is well warmed in these low latitudes; so much so, in fact, that the temperature increases inland even to considerable altitudes above sea level; and in this higher temperature of the land, the cool

ocean air departs farther and farther from its point of saturation. The conditions on the west coast of southern Africa are quite analogous. To a somewhat less marked degree, the phenomenon is repeated on the coast of California and on the west coast of northern Africa. All these coasts have a marked tendency toward a deficiency of rainfall so long as the cool current flows along them, and so long as there is cold water off-shore.

On the contrary, the coasts along which flow the warm equatorial branches of the oceanic circulation have abundant rainfall, while at the same time the winds, at least during the warmer season, also blow from lower to higher latitudes. The eastern coasts of the continents, therefore, have a plentiful supply of rain from the equator into the temperate zone. The whole eastern coast of Australia has abundant precipitation, as have the eastern coasts of south Africa and of South America, in contrast to the corresponding west coasts in the same latitudes. Eastern North America and eastern Asia also have enough rain. The proximity of a warm ocean increases the rainfall because the air which is nearly saturated at a high temperature, is frequently cooled, and the water vapour which it contains is condensed.

In higher latitudes, where the contrasts in the temperature conditions between continents and oceans are reversed from summer to winter, and where a warm ocean current flows along the coast, there is a tendency to a winter maximum of rainfall, because at this season the warm air from the ocean is greatly cooled over the cold land. In summer, on the other hand, when the land is warmer than the sea, the precipitation decreases both in frequency and in amount. This is the case on the northwestern coasts of Europe and of North America. It is true that the prevailing southwesterly and westerly winds on the eastern side of the low pressure areas which lie over the northern oceans tend to bring about this same result, and are indeed the cause of the warm ocean currents themselves. In the southern hemisphere, the South American continent is the only one which reaches far enough into the higher latitudes to have these abundant rains on its west coast. The Chilean coast south of latitude 40° S., where the prevailing winds are from west and northwest, is even more rainy than the coasts of Norway and of northwestern North America. It is a striking fact that these coasts are also peculiarly alike in that they rise abruptly from the sea, with their mountain ranges everywhere closely following the coast, and that their shore lines are indented with numerous fiords whereby the ocean gains access to the interior.

CHAPTER XI.

INFLUENCE OF FORESTS ON CLIMATE.

Influence of forests on temperature.—The influence of forests on climate is a question which has given rise to a great deal of discussion. We shall confine ourselves to a very brief consideration of the subject.[1]

It has already been clearly seen, in the cases of equatorial South America and Africa, that forests considerably reduce the mean temperature of the air, especially during the warmer portion of the year. Woeikof has shown that this same effect is probable in Assam and the Malay Archipelago.[2] The summer temperatures, even in the middle latitudes, are lowered by the influence of forests. Forests prevent the occurrence of high air temperatures by shading the ground, which may be heated to 60°-80° in dry regions; by increasing the surface from which radiation takes place; by the active radiating power of the leaves; by the increased evaporation from a large surface,[3] and the

[1] A useful summary of this question will be found in *Forest Influences*, Bulletin No. 7, U.S. Dept. of Agriculture, Forestry Division, 8vo, Washington, D.C., 1893, p. 197. This contains the following papers: "Introduction and Summary of Conclusions," by B. E. Fernow; "Review of Forest Meteorological Observations; A Study preliminary to the Discussion of the Relation of Forests to Climate," by M. W. Harrington; "Relation of Forests to Water Supplies," by B. E. Fernow; "Notes on the Sanitary Significance of Forests," by B. E. Fernow; "Determination of the true Amount of Precipitation and its Bearing on Theories of Forest Influences," by Cleveland Abbe; "Analysis of Rainfall with Relation to Surface Conditions," by George E. Curtis. See also a brief summary by J. Nisbet: "The Climatic and National Economic Influence of Forests," *Nature*, XLIX., 1893-94, 302-305.

[2] *Die Klimate der Erde*, Chap. XIII.; "Klimatologische Zeit- und Streitfragen," *M.Z.*, V., 1888, 191-195.

[3] An interesting result of this action has been deduced by Ney. He has attempted to explain the occurrence of heavy frosts by the fact that sensible heat is rendered latent in consequence of the sudden increase of evaporation when the

resulting reduction of temperature; and by the frequent production of fogs and clouds.[1]

Influence of forests on humidity.—Forests increase the relative humidity, and they decrease evaporation from the ground by the shade which they give, and by checking the movement of the air. They also increase the amount of water in the soil, notwithstanding the large supply of water which they need themselves, and act as regulators of the water circulating through the ground, and of the water supply of brooks and streams.[2] This regulating action on the part of the forests is most marked on all steep slopes. The water resulting from rainfall is retained within the forest, and is prevented from flowing off rapidly. A rapid flow-off is very undesirable. It is followed by a deficiency of water supply, and it does much damage by washing away the weathered upper surface of rocks, and the soil, thus leaving bare rocks, overloading the streams, and causing them suddenly to overflow their banks.[3]

The effect of the forest in holding the winter snow cover, and in favouring its slow melting in the spring, is of notable importance in many districts. This beneficial action is, however, accompanied by a

leaves appear in spring (C. E. Ney: "Der vegetative Wärmeverbrauch und sein Einfluss auf die Temperaturverhältnisse der Luft," *M.Z.*, II., 1885, 445-451). We are more inclined to attribute these frosts to the rapid increase in extent of a surface which is an exceptionally good radiator. For a discussion of the active radiating power of leaves, which equals that of soot, the reader is referred to A. G. Mayer: "Radiation and Absorption of Heat by Leaves," *Amer. Journ. Sci.*, 3d Series, CXLV., 1893, 340-346 (*M.Z.*, X., 1893, 319-320).

[1] The observations of Breitenlohner ("Wärmereflexion durch Laubwerk," *M.Z.*, X., 1893, 197-198), and Hartl, which were concerned with reflected heat, do not contradict this statement. It suffices to call attention to the fact, which may be observed almost any day, that some kinds of clouds dissolve over open country, and to the observations made by aëronauts, that balloons descend over forests, as they do over rivers and lakes, thus showing a tendency to a descending movement of the air.

[2] See Wiener: "Russische Forschungen auf dem Gebiete der Wasserfrage," *Forsch. auf dem Geb. der Agrikulturphysik*, XVIII., 1895, 433, etc.

[3] Measurements of the amount of rainfall which flows off on the surface in a forested and in an unforested area in Alsace, gave the relative danger of flood as 1 in the forested to 2 in the unforested area. (Jeandel, Cantégril et Bellaud: "Etudes expérimentales sur les Inondations," *Comptes Rendus*, LI., 1860, 1011-1015.) Excellent examples from the dry portions of India, quoted in *Nature*, XLI., 1889-90, 123, show the effects of deforestation in increasing floods and erosion, and in overloading the rivers of that region. Other cases are cited in Berthold Ribbentrop's *Forestry in British India*, Calcutta, 1900.

lowering of the spring temperature, and a slow rise of temperature during the spring months.[1]

Influence of forests on rainfall.—The question whether forests can increase the amount of rainfall, and the further question as to the amount of this increase, if there be any, cannot yet be definitely answered. In view of the influences of forests upon the other meteorological elements, which have already been noted, it may be concluded, with a good deal of certainty, that, so far at least as the tropics are concerned, forests may actually increase the amount of rainfall. The observations made under Blanford's supervision in the central provinces of India, support this view,[2] and Ribbentrop believes that the wholesale destruction of forests in India has had the most deteriorating effect on the climate, although he does not go so far as to maintain that reforestation might improve the climate to such an extent as to prevent the recurrence of droughts. Müttrich gives similar results for Germany, which seem to point toward an increase in the rainfall as a result of extended reforestation.[3] Hettner[4] also comes to the conclusion that a surface covering of vegetation influences precipitation. In the Cordillera at Bogotá, clouds with rain falling from them can be seen hanging over the forests, while near by, over ground which is covered with shrubs, or is used for agriculture, the sky is blue and the sun is shining. It appears, further, that this open country has been deforested, and that with the change in the covering of the soil, the climate has also changed to some extent.[5] On the other hand, in the case of North America, the question as to whether the increasing deforestation in the east, and the more extended cultivation of the soil in the west, have had any effect upon the amount of rainfall, has been investigated without leading to any decisive answer.

In general, the rainfall is to be looked upon as the cause, and the condition of the cover of vegetation as the effect.

[1] See also A. Müttrich: "Ueber den Einfluss des Waldes auf die Lufttemperatur nach den in Eberswalde an verschieden aufgestellten Thermometern gemachten Beobachtungen," *M.Z.*, XVII., 1900, 356-372; and P. Schubert: "Der Einfluss der Wälder auf das Klima," *ibid.*, 561-564.

[2] H. F. Blanford: "The Influence of Indian Forests on the Rainfall," *Journ. Asiat. Soc. Bengal*, LVI., Part II, 1887, 1-15 (*M.Z.*, V, 1888, 235-237).

[3] A. Müttrich: "Ueber den Einfluss des Waldes auf die Grösse der atmosphärischen Niederschläge," *Das Wetter*, IX., 1892, 46-48; 68-71; 90-96 (*M.Z.*, IX., 1892, 306-308).

[4] *Regenvertheilung, Pflanzendecke und Besiedelung der tropischen Anden*, Berlin, 1893.

[5] "Die Cordillere von Bogotá," *Pet. Mitt.*, Ergänzungsheft 104, 1892, 73-76.

Extended forests, even in middle and higher latitudes, certainly have some influence in increasing the frequency of rainfalls, but it is naturally almost impossible to determine the extent of this influence by observation and measurement. It must, nevertheless, be admitted that it was formerly the habit very greatly to overestimate this effect of forests in increasing rainfall, and the natural reaction has led to the present tendency altogether to deny such an influence. A surface which keeps the air moist and cool, and from which there is as great an evaporation as takes place from extended forests, must have a tendency to increase the amount and the frequency of precipitation, as contrasted with an open country which is dry, but over which conditions are otherwise similar. Small and scattering groves of trees will naturally not have this influence. It is, however, perfectly clear that the most important of all the effects of forests is seen in mountains, where they prevent the soil from being washed away, keep the water from flowing off too quickly, and check the silting and the rapid flooding of the rivers.[1]

In the case of fog and hoar frost, an increase in the amount of precipitation can be directly proved. During the prevalence of dense fogs, a light rain falls, drop by drop, under trees and in forests, and this rain wets the ground thoroughly, while in the open country the ground remains dry. A celebrated case of this sort on the island of Ascension has been described by Abbe. "The principal water supply for the garrison of this naval station is gathered several miles away, at the summit of Green Mountain, the upper part of which has always been green with verdure since the island was discovered; almost all of this water comes from slight showers and steady dripping of trees enveloped in cloud-fog on the windward side of the mountain. Every exposed object contributes its drip; these do not condense the water; they simply collect it mechanically after it has been condensed in the uprising cooling air. Whatever fog-drops are not thus collected sweep on over the mountain, and quickly evaporate again."[2]

The earlier growth of grass under trees in the spring has been attributed to the more abundant supply of moisture there. In India, during seasons of drought, the grass growing under the shade of the forests has often saved large numbers of cattle from death by starvation. In the case of a hoar frost, the twigs and branches of trees collect a very considerable

[1] At the Congress of French Geographers, held at Lyons in 1894, Quénot cited many cases in which deforestation in France had injured the habitability and the cultivation of valleys in mountainous regions.

[2] *Forest Influences*, 121.

amount of moisture, which is not so gathered where there are no trees. A hoar frost which occurred in Altenburg, Hungary, in December, 1860, was closely observed by Wilhelm, who computed the amount of frost collected on shrubs 1 to 2 m. high as equal to 1·9 mm. of rainfall in this single instance. Where there are tall trees, with many branches, the surface undoubtedly receives much more water, and when this process of collecting water vapour from the air is frequently repeated, as happens in some winters and in certain districts, the surface can, in this way, receive a considerable supply of water.[1]

Breitenlohner has made similar observations in the Vienna forest;[2] and Fischbach notes that he has frequently observed that hoar frost shaken down from the trees by the wind, has made it possible to haul lumber on sledges in the Black Forest during winters when there was little snowfall.[3] This is, therefore, a case of a direct increase in the amount of precipitation resulting from the presence of a forest, because the vapour either does not collect at all over open fields, or else collects in very small quantities only. The forest alone, by means of its branches and leaves, can effectively gather the water which is suspended in the air during fogs, and can supply it to the earth's surface.

The protection afforded by forests against strong winds is another important function performed by groups of trees. This effect consists not only in a decrease in the velocity of these winds within the forest, but also in its vicinity. Forests thus check the rapid drying of the soil, and, in winter, the drifting of snow. On the other hand, the diminished air movement does favour the occurrence of spring and autumn frosts. The almost complete disappearance of peaches in the state of Michigan has been ascribed to deforestation, as a result of which the cold north and northwest winds are supposed to be stronger and much more injurious than they used to be. Indeed, Curtis recommends belts of trees as the best protection against cold, as well as against hot or dry winds. Forests decrease the violence of the winds, and diminish the injurious effects which they may produce by their severe cold, heat or dryness. In the year 1888, 21 million bushels of

[1] J. Lamont: "Einfluss der Feuchtigkeit auf die Temperatur der freien Luft," *Z.f.M.*, II., 1867, 126-127.

[2] *Forsch. auf dem Geb. der Agrikulturphysik*, II., 497.

[3] K. von Fischbach: "Ueber den Einfluss des Waldes auf atmosphärischen Niederschlag und das Eindringen des Wassers in den Boden," *M.Z.*, X., 1893, 194-196.

corn were lost in the state of Kansas, through the influence of hot dry winds.[1]

In general it may be said that the presence of extended forests in continental interiors has an influence upon the climate, especially in summer, which to a very slight degree gives it the characteristics of a coast climate. Even the presence of extended swamps and fens has a considerable climatic influence, for these surfaces have a cooling effect upon the air, and increase its humidity. The surface temperatures under these conditions are low, and the heat of the day and of the summer penetrates but slowly, and to a slight depth, into a swampy soil. Hence the ground remains frozen for a long time in such places, and swamps are therefore one of the chief reasons why the soil in the far north is continuously frozen.[2]

[1] G. E. Curtis: "Winds Injurious to Vegetation and Crops," Bulletin 11, U. S. Weather Bureau, Part II., 1895, 435-444. See also I. M. Cline: "Summer Hot Winds on the Great Plains," Bull. Philosoph. Soc., Washington, XII., 1894, 335-348; and *Amer. Met. Journ.*, XI., 1894-95, 175-186.

[2] There are some instructive diagrams showing the diurnal march of temperature in a swampy surface as compared with an open surface in a report by Th. Homén: "Bodenphysikalische und Meteorologische Beobachtungen," Berlin, 1894.

CHAPTER XII.

MEAN TEMPERATURES OF PARALLELS OF LATITUDE AND OF THE HEMISPHERES.

Mean temperatures of parallels of latitude.—The temperature of the air at the earth's surface depends primarily, as has already been explained, upon latitude. It is, however, also directly dependent upon the fact whether the earth's surface at any given latitude is composed of land or water. A hemisphere composed wholly of land or wholly of water would present the simplest condition. As a matter of fact, it is in the southern hemisphere alone that the distribution of temperature approaches that which would prevail over a water hemisphere. The northern hemisphere, although it contains the largest continental areas, is nevertheless far from being a true land hemisphere, because its surface consists of only 40 per cent. of land and 60 per cent. of water. There is, moreover, no single parallel of latitude, the whole of whose course lies only over the land. Hence the conclusions, which may be drawn with more or less accuracy from the dependence of observed temperatures upon the distribution of land and water, afford us our only basis for determining the distribution of temperature over a hemisphere wholly composed of land.

We owe the first serious inquiry along these lines to the English physicist, James D. Forbes.[1] The basis of the investigations carried on by Forbes, and in later years by Spitaler, were the so-called *normal* temperatures, first computed by Dove for every ten degrees

[1] J. D. Forbes: "Inquiries about Terrestrial Temperatures," *Trans. Roy. Soc.*, Edinburgh, XXII., Pt. I., 1859, 75-92.

of latitude in their relation to the percentage of land and water along each parallel. These temperatures must therefore be considered first.

In his epoch-making work, *The Distribution of Temperature over the Earth's Surface*, published in Berlin in 1852, Dove computed the mean temperatures of the parallels of latitude at intervals of ten latitude degrees for each of the twelve months, and for the year. This was done by taking from the monthly isothermal charts the temperatures at thirty-six equidistant points along any given latitude circle (*i.e.*, at every ten degrees of longitude) and deriving the mean from these thirty-six different temperatures. This mean temperature is also regarded as the *normal temperature* of the given parallel of latitude, although the term is not exact. This mean, however, depends not upon the latitude alone, but also upon the relation of land and water along the given parallel, and this relation changes from one parallel to another. After Dove, Spitaler,[1] and more recently Batchelder,[2] determined the mean temperatures of the parallels of latitude for January, July, and for the year. Spitaler used, as the basis of his work, the author's isothermal charts in Berghaus's new physical atlas (Gotha, 1887), and Batchelder used the isothermal charts prepared by Buchan, and published in the *Report on Physics and Chemistry*, Vol. II. of the *Challenger* expedition.[3]

In view of the importance of these data for climatology in general, and especially for the discussion which follows, they are included here. The second and third columns of the table (a and b) give the relative distribution of land and water along the different parallels of latitude. These values are taken from Penck's *Morphologie der Erdoberfläche*. The figures under b are determined by taking into account the parallels 5° away on either side. Thus, for example, for latitude 60° we have

$$\tfrac{1}{2}[60 + (65 + 55) \div 2].$$

The values given under b are the best for use in discussing the relation between the temperature and the amount of land surface along any parallel.

[1] R. Spitaler: "Die Wärmevertheilung auf der Erdoberfläche, *D. W. A.*, LI., 1886, Pt. II., 1-20.

[2] S. F. Batchelder: "A New Series of Isanomalous Temperature Charts," *Amer. Met. Jour.*, X., March, 1894, 451-474. These charts are also published in Bartholomew's *Atlas of Meteorology*, Plate 2.

[3] See also *M. Z.*, XVII., 1900, 36-39.

MEAN TEMPERATURES OF THE PARALLELS OF LATITUDE.

Latitude.	Amount of Land Surface, in per cents.		Mean Annual Temperature.		January.	July.	Difference.
	a.	b.	Spitaler.	Batchelder.	Mean, Spitaler and Batchelder.		
			°	°	°	°	°
N. Pole, -	—	—	−20·0	−20·0	(−38·0)	(0·0)	38·0
80°, - -	22	24	−16·5	−16·9	−33·5	1·8	35·3
70°, - -	55	54	−9·9	−10·2	−26·0	7·0	33·0
60°, - -	61	64	−0·8	−1·2	−15·8	14·0	29·8
50°, - -	56	55	5·6	5·8	−7·0	18·1	25·1
40°, - -	46	47	14·0	13·9	4·9	24·0	19·1
30°, - -	43	42	20·3	20·2	14·6	27·3	12·7
20°, - -	33	32	25·6	24·9	21·9	28·3	6·4
10°, - -	24	24	26·4	27·1	25·8	26·9	1·1
Equator, -	22	23	25·9	26·6	26·4	25·6	0·8
10°, - -	20	23	25·0	25·7	26·3	23·9	2·4
20°, - -	24	23	22·7	23·3	25·4	20·0	5·4
30°, - -	20	18	18·5	18·3	21·8	14·6	7·2
40°, - -	4	5	11·8	12·2	15·6	9·0	6.6
50°, - -	2	2	5·9	5·3	8·3	2·9	5·4
60°, - -	0	1	(0·2)	−1·1	1·6	(−3·8)	—

Position of heat equator.—It appears from the above table that the highest mean annual temperature is found at latitude 10° N. This parallel is therefore the *thermal*, or *heat equator*. The equator is the warmest parallel only during the winter of the northern hemisphere, in January; while in July the highest temperature is seen to be somewhat north of latitude 20° N. This displacement of the heat equator into the northern hemisphere shows very strikingly the influence of the greater extent of the land surface in that hemisphere, for land areas in low latitudes are warmer than the ocean. The movement, from the southern into the northern hemisphere, of a considerable body of warm water, under the influence of the trade winds, to which reference has already been made (see page 188), is another factor in causing this displacement. The northern portion of the Indian Ocean, which is well surrounded by land, and the archipelago southeast of Asia and north-east of Australia, in many parts of which the water is comparatively shallow, are likewise sources of warmth which are not present in the southern hemisphere. The oceans of the northern hemisphere, which

are either wholly or partially shut off from the Arctic Ocean and its ice fields, are, as a whole, warmer than the oceans south of the equator. It should, however, be noted that the greater land mass of the northern hemisphere is effective in thus enclosing these northern oceans.

Temperatures of northern and southern hemispheres.—The average amount of land in the western, or water, hemisphere, between longitude 20° W. and 160° E., is only about half as large between latitudes 5° and 25° in the northern hemisphere as in corresponding latitudes of the southern hemisphere, yet the mean temperature of the western hemisphere, between latitudes 5° and 25° N., is 24·6°; while in the same latitudes of the southern hemisphere it is only 23·3°. The southern latitudes are therefore 1·3° cooler. Further, although there is only 6 per cent. of land at latitude 15° N. in the western hemisphere, while in latitude 15° S. there is 20 per cent., yet the temperature in the northern latitudes is still 1·3° higher than in the southern. Thus, not only the land, but also the water, is warmer in these latitudes of the northern hemisphere.[1]

Mean temperatures of the earth as a whole in different months.—Dove first called attention to the notable fact that the mean temperature of the earth as a whole does not remain constant throughout the year, as should theoretically be the case in view of the conditions of insolation, but that it rises from January to July, and that the temperatures of the northern hemisphere are therefore the controlling ones for the earth as a whole. According to Spitaler, these mean temperatures are as follows:—

MEAN TEMPERATURES OF THE HEMISPHERES IN JANUARY AND JULY.

	January.	July.	Difference.	Mean.
	°	°	°	°
Northern Hemisphere,	8·0	22·5	14·5	15·2
Southern Hemisphere,	17·5	12·4	5·1	14·9
Earth as a whole, -	12·7	17·4	4·7	15·0

The temperature of the earth as a whole, therefore, increases nearly 5° from January to July, in consequence of the high July temperature of the northern hemisphere which comes simultaneously with the milder

[1] A. Woeikof: "Bemerkungen ueber den Einfluss von Land und Meer auf die Lufttemperatur," *M.Z.*, V., 1888, 17-21. The means given in the text have been determined without reference to the unequal circumference of the parallels of latitude; a fact which, however, does not vitiate the comparison.

winter temperature of the southern hemisphere. On the other hand, the low summer temperature of the latter corresponds with the low January temperature of the former. The northern land hemisphere, if the expression is allowable, has a cold winter and a hot summer. The annual range of temperature is 14·5°. The southern water hemisphere has a cool summer and a mild winter, the annual range of temperature being but a little more than one-third as great as that of the northern hemisphere. The northern hemisphere has a moderate continental climate, its surface being about 40 per cent. land. The southern hemisphere, with only 17 per cent. of land, has a fairly typical marine climate.

The mean temperature of the earth as a whole is about 15°, both hemispheres having nearly the same temperature. The southern hemisphere is, however, probably somewhat cooler, chiefly in consequence of the heat which is carried across the equator by the strong southeast trade drift.[1]

[1] The mean temperatures of both hemispheres as computed by different investigators are as follows:—

Northern hemisphere: Dove, 15·5°; Schoch, 15·1°; Ferrel, 15·3°; Spitaler, 15·4°.

Southern hemisphere: Sartorius von Waltershausen, 15·8°; Schoch, 14·9°; Ferrel, 16·0°; Hann (1882), 15·4°; Spitaler, 14·8°; Hann (1896), 14·7°. (W. Schoch: *Darstellung der mittleren Jahrestemperatur als Funktion der geographischen Breite und Länge*, Zürich, 1856. Based on Dove's isothermal charts. William Ferrel: *Meteorological Researches for the Use of the Coast Pilot*, I., U.S. Coast and Geod. Surv., Washington, 1877. Based on Buchan's original isotherms. J. Hann: "Ueber die Temperatur der südlichen Hemisphäre," *S. W. A.*, LXXXV., 1882, Pt. II., 6-29).

It seems well to include here a purely theoretical computation of the mean temperature of the earth's surface, and to compare this with the results of observation. Christiansen has made the following calculation. It is assumed that, of the radiant energy which comes to the earth from the sun, 2 calories (per sq. cm. per minute) are absorbed and applied to raising the temperature. This assumption is consistent with the results of observation. (Langley's solar constant is 3, of which value about one-half, *i.e.*, 1·5, reaches the earth's surface with a zenithal sun, and the diffuse radiation from the atmosphere can easily supply enough to make 2 effective calories.) If this assumption is made, the amount of heat effective in warming the earth, received by the latter, is $2 \div 4$. The whole surface (S) receives the heat of a pencil of radiation whose cross section at the earth is a great circle of the earth. If we take Stefan's law of radiation, and if the temperature of the earth is t, the amount of heat radiated from the earth is $a\,(273+t)^4S$, in which a is the coefficient 0·0000000000728. Since the temperature of the earth is constant, the gain and loss of heat must be equal, and we have

$$0{\cdot}728 \times 10^{-10}(273+t)^4 = 2 \div 4,$$

which gives

$$t = 15^\circ$$

This law of radiation therefore gives the actual mean temperature of the earth's surface. Conversely, as this same mean temperature is obtained by observation, this computation may be taken as proof that the effective radiation from the zenithal

Mean temperatures of the hemispheres.—The unequal distribution of land and water, and the permanent currents in atmosphere and ocean, also cause differences of temperature from east to west along the same parallel of latitude. Spitaler obtains the following values for the mean temperatures of an eastern and a western hemisphere, the two hemispheres being divided by the meridians of 80° W. and 100° E., on the basis of the distribution of the intensity of terrestrial magnetism.

MEAN TEMPERATURES OF EASTERN AND WESTERN HEMISPHERES.

	Eastern Hemisphere.	Western Hemisphere.
	°	°
North, - - - - -	16·7	13·9
South, - - - - -	14·3	14·9
Earth as a whole, - -	15·5	14·4

The eastern hemisphere has a mean temperature of 9·4° in January, and 23·6° in July. The western hemisphere has 6·5° in January, and 21·5° in July, and is therefore cooler in both seasons. Supan has obtained the mean temperatures of both hemispheres, divided, in accordance with the usual custom, by longitude 20° W., and 160° E.[1]

sun must be very nearly 2 calories. Actually, the effective radiation is less than 2, because of the obliquity of the mean path, and a must be less than for the ideal black body.

Christiansen also computed the mean temperatures of the most important parallels of latitude in the same way, and obtained the following results :—

	Equator.	Tropic.	45°.	Polar Circle.	Pole.
	°	°	°	°	°
Computed, - - -	29·5	24·0	7·5	−18·5	−30·0
Observed, - - -	26·2	23·3	9·3	(−4·0)	(−20·0)

The agreement is quite close, for, in consequence of the diffusion of heat by air and ocean currents, the lower latitudes must have lower temperatures, and the high latitudes must have higher temperatures, than would be obtained by means of the mathematical formula. The greatest positive departure is to be expected near the Arctic and the Antarctic circles. (*Dansk Vidensk. Selsk. Forh.*, Copenhagen, 1886, 85-108).

[1] A. Supan: Review of Spitaler's *Die Wärmevertheilung auf der Erdoberfläche. Pet. Mitt.*, XXXIII., 1887. *Litteraturber.*, 90-91. The western hemisphere has 17 per cent. land and 83 per cent. water; the eastern has 37 per cent. land and 63 per cent. water (from 80° N. to 70° S.).

By Supan's determination, the western water hemisphere, between latitude 20° and latitude 60° N., is warmer in January than the eastern land hemisphere. At latitude 50° N. the temperature is 8·7° higher in the former than in the latter. In July, on the other hand, the latter is warmer from latitude 70° N. to the equator; and nearly up to latitude 20° N. it is almost 5° warmer. At latitude 20° N. the eastern hemisphere has a temperature of 30·1°; the western, 26°.

Mean temperatures of meridians.—Buys-Ballot also determined the mean temperatures of the meridians for intervals of 5°, for January, July, and the year; the area embraced in this investigation being that between latitude 80° N. and 55° S. These figures show the existence of considerable differences of temperature, two maxima and two minima being noticeable in all seasons, which is quite in accordance with the prevailing longitudinal distribution of land and water.[1]

In January, the warmest meridians are 160° to 170° W., with a temperature of 10·5°; and 10° to 35° E., with 12·5°. The coldest meridians are 95° to 105° W., with 5·4°; and 100° to 115° E., with 2·8°. The maximum difference is 9·7°. In July, the warmest meridians are 100° to 115° W., with 18·3°; and 120° to 115° E., with 19·2°: the coldest are 175° to 180° W., with 16·0°; and then 125° to 130° W., and 15° to 20° W., with 16·1°. The maximum difference is only 3·2°.

In the northern hemisphere, the meridian of 120° E. probably has the most strikingly continental type of temperature conditions. The greatest contrast with the conditions along this meridian is probably to be found along the meridian of 30° W. The mean temperatures along these two meridians have been determined by the author, on the basis of Batchelder's data, with the following result:—

MEAN TEMPERATURES ALONG THE MERIDIANS OF 120° E. AND 20° W.

Equator to 80° N.	January.	July.	Year.
	°	°	°
120° E., - - -	−6·6	21·1	7·4 Land.
20° W., - - -	10·7	17·3	12·8 Water.

South of latitude 20°, there is hardly any difference of temperature between east and west. If, therefore, the extra-tropical latitudes alone are taken from 20° to 80° N., we have the following:—

[1] *Verdeeling der Warmte over de Aarde.* Amsterdam, 1888.

MEAN TEMPERATURES ALONG THE MERIDIANS OF 120° E. AND 20° W. BETWEEN LATITUDES 20° AND 80° N.

Latitudes 20°-80° N.	January.	July.	Year.
	°	°	°
120° E., - - -	−15·9	19·4	1·7 Land.
20° W., - - -	6·3	14·6	8·7 Water.
Difference, - -	+22·2	−4·8	+7·0

These are probably the greatest contrasts of temperature between east and west. The temperatures given in the above table are the means of longitudes 110°, 120° and 130° E., and 20°, 30° and 40° W.

The mean temperatures of all oceans and continents between latitudes 90° N. and 50° S. have been calculated by von Tillo, with the following results:—

	January.	July.	Year.	Range.
	°	°	°	°
Mean Temperature of Oceans, -	17·9	19·2	18·3	1·3
,, ,, Continents, -	7·3	22·9	15·0	15·6

The oceans are 3·3° warmer than the continents. In the January and the July temperatures, the northern hemisphere turns the scale, as may be clearly seen. The range of temperature over the continents from January to July is twelve times as great as over the oceans.

Mean temperatures of land and water hemispheres.—These very general results concerning the distribution of temperature over the surface of the earth may properly be supplemented by the results obtained by Forbes,[1] in his studies of the characteristics of typical continental and marine climates. In connection with this, reference may be made to the table given on page 200, which may serve as a basis for our consideration.

The mean temperature of a parallel of latitude may, as has already been seen, be considered as dependent upon the latitude itself, and upon the relative proportion of land and water along that parallel. The transfer of heat, by air and ocean currents, has but little effect upon this temperature, because it may be assumed that the warm currents along one meridian are very largely compensated by the returning

[1] *loc. cit.*

cold currents along another meridian. A formula might therefore be devised which should express the temperature of every parallel of latitude as a function of the latitude and of the distribution of land and water along that parallel. Such a formula was actually worked out by Forbes, and is given in the Appendix to this chapter. This formula, although derived from observations made in the northern hemisphere only, nevertheless satisfactorily represents the distribution of temperature in the southern hemisphere as well, notwithstanding the great differences in the distribution of land and water in corresponding latitudes. This fact may be taken as evidence that it accomplishes somewhat more than an interpolation formula.

The formula given by Forbes makes it possible to reach some approximation concerning the temperatures over a hemisphere composed wholly of water, or wholly of land. Thus we obtain the following table :—

TEMPERATURES AT EQUATOR AND AT POLE IN WATER AND LAND HEMISPHERES.

	Water Hemisphere.	Land Hemisphere.
	°	°
Temperature at Equator, - -	22·1	43·2
,, Pole, - - -	-10·8	-32·0

If it be allowable, in the light of what has been said, to regard the temperature of a parallel of latitude as the resultant of the mean temperatures over its land and over its water areas, then the mean temperatures of a water hemisphere and of a land hemisphere may be computed in still another way. For each parallel of latitude, two equations may be constructed, one for the northern, and one for the southern hemisphere. These equations express the temperature of that parallel as the resultant of the unknown mean temperatures of the known areas of land and of water. By eliminating the two unknown terms from each of the two equations for each parallel, we obtain the mean temperature of a parallel running over a perfectly uniform surface of land and water.[1] By this method, Forbes obtained the temperatures at latitudes 10°, 20°, 30°, and 40°, for a land and a water hemisphere, and these he represented graphically. A slight extension, drawn by free hand, of the two symmetrical branches of the curve, then gives the temperature at the equator, as well as the

[1] See *Appendix* for an illustration of this method.

latitude at which the two curves intersect, *i.e.*, the latitude at which the temperature of the water hemisphere becomes the same as that of the land hemisphere. The temperature at the equator is found to be as follows:—

Surface wholly water, 22·1°.

Surface wholly land, 46·3°.

The difference is 24·2°. This result shows a perfectly satisfactory agreement with the values previously obtained by the general formula. The mean temperatures of a land and water hemisphere, as obtained by both these methods, namely, by formula and by graphic representation, are as follows:—

MEAN TEMPERATURES OF LAND AND WATER HEMISPHERES.

	0°	10°	20°	30°	40°	50°
	°	°	°	°	°	°
Land Hemisphere, - -	44·8	42·5	36·4	26·0	15·7	3·6
Water Hemisphere, -	22·2	21·2	19·6	17·4	12·7	7·6
Difference, - - -	22·6	21·3	16·8	8·6	3·0	−4·0

Between latitude 40° and latitude 50° is the parallel on which both water and land hemispheres have the same temperature. In higher latitudes, the water hemisphere is warmer than the land hemisphere. Even Dove's isothermal charts of the northern hemisphere show that the transition takes place between latitudes 40° and 45°, and the graphic representation, above referred to, shows more precisely that this transition occurs at latitude 42½°. Beyond this latitude, as far as the pole, a water hemisphere is warmer than a hemisphere wholly covered by land.

Madsen[1] has recently attempted to express, by means of a series of equations, the observed mean temperatures of the earth's surface as a function of the geographical coordinates, latitude, longitude and altitude. For a discussion of this subject the reader is referred to the original monograph.

On the basis of the mean temperatures of the parallels of latitude, as determined by himself, and of the distribution of land and water, as determined by Dove, Spitaler obtained another formula, which

[1] G. J. Madsen: *Thermo-Geographical Studies. General Exposition of the Analytical Method applied to Researches on Temperature and Climate.* Copenhagen, 1897.

expresses the temperature of each parallel of latitude as a function of the latitude, and of the relations of land and water along that parallel.[1]

The formula leads to the following results :—

MEAN TEMPERATURES OF LAND AND WATER HEMISPHERES (SPITALER).

	Water Hemisphere.	Land Hemisphere.	Difference.
	°	°	°
Temperature at the Equator, - -	22·2	41·5	+19·3
,, ,, Pole, - - -	−9·5	−28·8	−19·3
,, of the whole Hemisphere,	13·8	20·2	+6·4

The highest temperature would be found in a hemisphere covered with land from the equator to latitude 45°, and with water from there to the pole. The lowest temperature would be found under the opposite conditions of distribution of land and water.

Spitaler worked out the mean temperatures of the earth under these conditions, and found that with a land surface from the equator to latitude 45°, and a water surface from there to the pole, the mean temperature of the earth would be 22·8°; while with a water surface from the equator to latitude 45°, and a land surface from there to the pole, the mean temperature would be 11·1°. The zone from the equator to latitude 45° in a water hemisphere would have a mean temperature of 18·2°; while in a land hemisphere the temperature would be 31·1°. The polar cap from latitude 45° to the pole in a water hemisphere would have 2·7°, and in a land hemisphere, −6·2°.

Such results as the foregoing do not give a very faithful representation of existing conditions. Nevertheless, they have some interest, for the reason that they give us a clearer idea than we should otherwise have of the influence of land and water upon the temperature of the earth's surface, as well as of the changes which would take place if the distribution of land and water on the earth's surface should vary. Von Kerner has actually made an interesting attempt to determine the

[1] This formula is as follows, when ϕ = the latitude, and n = the percentage of land along this latitude circle:

$$t\phi = -2{\cdot}43 + 17{\cdot}6 \cos\phi + 7{\cdot}1 \cos 2\phi + 19{\cdot}3 n \cos 2\phi.$$

If n in the above formula = 0, the expression holds true for a water hemisphere; and if $n = 1$, it is true for a land hemisphere.

mean temperature of the earth in Jurassic time, on the assumption that the distribution of land and water, at that epoch, was like that shown on Neumayer's map of the earth in the Jurassic. Von Kerner obtained from Neumayer's map the relative amounts of land and water[1] along the different parallels of latitude, and, from Spitaler's formula, the temperatures of a land and water hemisphere. It appeared from this calculation that, in Jurassic time, the temperatures from latitude 20° N. to latitude 40° S. were considerably higher than at present; that they were 6·5° higher along the equator, and that they were also higher then between latitude 50° and latitude 70° N. than now. At latitude 30° N., the temperatures were lower in Jurassic time than they are at present. The mean temperature of the southern hemisphere appears, if we accept Neumayer's distribution of land and water, to have been about 1·5° higher than that of the northern hemisphere, and the mean temperature of the earth as a whole appears to have been more than 2° higher than at present.[2]

Polar temperatures of land and water hemispheres.—Woeikof has raised certain valid objections against the polar and equatorial temperatures of a land and a water hemisphere which are obtained by the empirical formulae of Forbes and Spitaler.[3] It is difficult to see what causes the low temperature of 22° on the equator of a water hemisphere, since we have nothing approaching it even in the middle of the Pacific Ocean. Woeikof is of the opinion that the mean air temperature at the equator of a water hemisphere would be 26°, a result which is the same as that reached by the author.[4] Again, the temperature of 44·8° at the equator of an earth wholly covered with land, is probably too high.

When the temperature at the pole of a water hemisphere is found to be − 9° to − 10°, it is clear that we are already dealing with purely imaginary temperatures, because even salt water cannot cool below − 3° or − 4° without freezing, and the mean annual temperature over a surface of salt water should hardly be assumed to be lower than

[1] n, in the formula of Spitaler.

[2] F. von Kerner: "Eine paläoklimatologische Studie," *S. W. A.*, CIV., IIa., 1895, 286-291. Precht has recently used a constant n in the formulae of Forbes and Spitaler, for he assumes that land and water are everywhere uniformly distributed along all parallels of latitude. The value of n is then 24·4 per cent. By this method he has worked out new *normal temperatures*, to which, however, no more extended consideration can here be given. (W. Precht: "Neue Normaltemperaturen," *M. Z.*, IX., 1894, 81-90.)

[3] *Die Klimate der Erde*, I., 332.

[4] J. Hann: "Temperatur der südlichen Hemisphäre," *loc. cit.*

$-5°$ to $-6°$. If the water surface freezes, then we have to do with a polar temperature similar to that in the case of a land surface. Notwithstanding these difficulties, however, such calculations as these are of importance in the study of theoretical climatology, because they furnish us with some basis for determining the influence of the south polar ice cap in lowering the temperature in that region.

As regards the polar temperature of a land hemisphere, Woeikof rightly calls attention to the fact that this could not be so low as that obtained above, because there would be no snow on a land hemisphere, and therefore the summer temperature at the pole would have to be very high. Following Woeikof's suggestion, we may conclude that the January temperature could hardly fall below $-50°$, especially if the land were level[1] and the July temperature over a dry land area without a snow cover could easily rise to 16° to 20°, if not higher. Under these conditions we should have, at the pole, a mean annual temperature of between $-18°$ and $-15°$, or even higher still. It is naturally most difficult to estimate the temperatures over a land hemisphere, because the actual conditions of the earth's surface are so different from the hypothetical ones.

Zenker's normal temperatures of land and water hemispheres.—Zenker has determined the normal temperatures of a hemisphere composed wholly of land, or wholly of water, by a method entirely different from that used by Forbes and Spitaler. By a purely theoretical process, similar to that adopted by Angot, he first calculates the amounts of heat received on the earth's surface at every parallel of latitude, taking into account, not only the absorption of solar radiation by the atmosphere, but also the diffraction in the atmosphere, and the reflection from land, water, and snow surfaces. The diffuse radiation is thus also taken into consideration. Zenker believes that his estimates of the amounts of heat, which are naturally computed for a clear sky, nevertheless hold good in the case of nature, because clouds check radiation at night, just as they diminish insolation by day. In his last great work Zenker has, however, endeavoured to take account of the effect of cloudiness as well, and he has deduced certain factors for use in correcting for the different degrees of cloudiness.

Zenker believes that it may be assumed that the variation in radiation at different points is proportional to the differences in temperature. If there are two points at different latitudes which may be considered, in so far as their mean temperatures are concerned, to have a typical

[1] The air currents would hardly allow so low a mean temperature to be reached.

marine climate, and two other points which may be considered to have a typical continental climate, we are in a position to determine the normal temperature of a parallel of latitude in a purely marine and in a purely continental climate. This is the starting point of Zenker's investigation. The typical marine climate Zenker very properly seeks over the Pacific Ocean in the southern hemisphere. There, at latitude 20° S., he finds the isotherm of 23°, and at latitude 50° S. the isotherm of 8°. The difference in insolation between latitudes 20° and 50° is 801 units, according to Zenker's table as given in his latest publication.[1]

Now, as this difference in the amount of total radiation corresponds to a difference of temperature of 15°, the unit of the difference in the amount of total radiation is 0·0187°. The further determination is simple. As the computed difference of heat energy received between latitude 20° and the equator is 166 (see table in foot-note), the difference of temperature is 3·1°, and the temperature at the equator in a marine climate is therefore 23·0° + 3·1° = 26·1°, which is a very close agreement with the results previously obtained. Similarly, the normal temperatures of the remaining parallels of latitude are obtained.[2]

It is evident that in this method of determining the normal temperatures the most important thing is to obtain the correct basal temperature for the reduction of the amounts of heat to degrees of temperature.

[1] TOTAL ANNUAL RADIATION RECEIVED BY THE EARTH'S SURFACE (ZENKER).

Latitude.	Land.	Water.	Latitude.	Land.	Water.
0°	2591	2528	50°	1654	1561
10	2550	2485	60	1314	1218
20	2429	2362	70	1038	935
30	2233	2158	80	903	785
40	1969	1887	90	882	761

These values are relative, and are to be read as units of the fourth decimal place. They are based upon the unit of Wiener, *i.e.*, that quantity of radiation which the unit of surface would receive at the equator if the sun were to remain in the zenith, and at its mean distance throughout the whole year.

[2] From the given temperature and the amount of radiation received at the equator, as well as the amount of radiation which corresponds to a difference of temperature of 1° (in the case we have taken this is 53·4; or, as it is now more accurately assumed by Zenker, 50·7), we obtain the equation

$$t = (\text{amount of solar radiation} \div 50{\cdot}7) - 23{\cdot}7.$$

This formula is the most convenient to use in the computation.

This process is clearly a somewhat arbitrary one. Nevertheless, there can be no doubt that this method of calculating the normal temperature of a typical marine and of a typical continental climate is less empirical than the method adopted by Forbes, and it furnishes a basis which is theoretically a better one for judging typical marine and continental climates than that obtained by the formulas previously given.[1]

It is much more difficult to determine the requisite values to be used in reducing the actual amounts of heat energy received to temperatures in the case of a continental climate, because there is no typical continental climate in the northern hemisphere over a sufficiently extended area to be available as a criterion in this connection. In his first publication, Zenker took a mean difference of temperature of 0·0385° as the unit difference in radiation. Starting with the isotherm of 0·5°, at latitude 50° N. in eastern Asia, we find the temperature at the equator in a continental climate 36·6°, the difference in radiation being 937 units. In his most recent extended investigation, Zenker finds a temperature at the equator of only 34·6°. This new method, which seeks to take the loss of heat by radiation into account, cannot be further discussed here.

NORMAL TEMPERATURES OF CONTINENTAL AND MARINE CLIMATES, AND OF THE SOUTHERN HEMISPHERE.

Normal Temperatures (Zenker).				Temperature of the Southern Hemisphere.	
Latitude.	In Continental Climate.	In Marine Climate.	Difference.	Computed.	Observed.
° 0	° 34·6	° 26·1	° −8·5	° 26·0	° 26·2
10	33·5	25·3	−8·2	25·5	25·4
20	30·0	22·7	−7·3	22·8	23·0
30	24·1	18·8	−5·3	18·1	18·4
40	15·7	13·4	−2·3	12·0	12·0
50	5·0	7·1	2·1	5·5	5·6
60	−7·7	0·3	8·0	−0·7	−1·0
70	−19·0	−5·2	13·8	−5·8	—
80	−24·9	−8·2	16·7	−9·1	—
Pole,	−26·1	−8·7	17·4	−11·3	—

[1] It is true that Zenker entirely disregarded the important factor of loss of heat by radiation in his first publication, but in his latest complete discussion he takes it into account [see *M.Z.*, XIII., 1896, (25)-(28)].

The normal temperatures are those given by Zenker in his latest publication.[1] The observed temperatures in the southern hemisphere are the means of the values obtained by Spitaler and Batchelder. The computed temperatures in the southern hemisphere were obtained by an interpolation formula derived by the author.[2]

It will be observed with satisfaction that the temperatures of a water hemisphere, which were calculated by Zenker from the amounts of radiation alone, agree very closely with the observed temperatures of the southern hemisphere. In the middle latitudes alone does Zenker's simple formula give temperatures 1·5° too high. We are, therefore, pretty clear as regards the distribution of the annual temperatures over a water hemisphere. There is no way of proving the accuracy of the computed temperatures of a land hemisphere. The temperature of the equator, as assumed by Zenker, is too low rather than too high, because annual means of 30° are, as a matter of fact, found as far north as latitude 15° N. The temperature of the pole is probably too low, because the influence of convectional currents, which would have to be a warming one in the circumpolar region, had to be left out of account

[1] W. Zenker: *Die Vertheilung der Wärme auf der Erdoberfläche*, Berlin, 1888, 62, *et sqq.*

"Ueber den klimatischen Wärmewerth der Sonnenstrahlen und ueber die zum thermischen Aufbau der Klimate mitwirkenden Ursachen," *M.Z.*, IX., 1892, 336-344, 380-394; X., 1893, 340-342.

"Die gesetzmässige Vertheilung der Lufttemperatur ueber dem Meere." With a chart, *Pet. Mitth.*, XXXIX, 1893, 39-44 (the chart shows the departures of the air temperature over the ocean from the theoretical values in a typical marine climate).

"Der thermische Aufbau der Klimate aus den Wärmewirkungen der Sonnenstrahlung und des Erdinnern," *Nova Acta Acad. Caes. Leop.-Carol. Germ. Nat. Cur.* (Halle), 1896, 1-252. (This great work embraces 252 quarto pages, and is accompanied by five charts).

[2] The author has determined the temperature at the equator in a typical marine climate to be 26·0°, on the basis of the mean temperatures of islands lying near the equator. As the mean temperature of the ocean on the equator is about 27°, and the air over tropical oceans is, according to Schott, about 0·8° cooler, this value may be considered to be accurate. For latitude 65° S., the author, following the observations of Ross and Borchgrevink, has inserted the temperature of −3·5°. Seven equations of condition gave

$$T_\phi = 26{\cdot}0 + 4{\cdot}54 \sin\phi - 40{\cdot}81 \sin^2\phi.$$

This equation gives the observed mean temperatures very accurately. Thus, for example, for latitude 65° the formula gives −3·4°. This formula gives the mean temperature of the southern hemisphere as 14·7°, this being a lower temperature than that originally obtained by the author, but in entire agreement with that reached by Spitaler.

altogether in Zenker's calculation. The influence of such currents must be very considerable in a continental climate in winter, because of the poleward decrease of temperature. The calculation must, therefore, inevitably result in giving too low polar temperatures. The normal temperatures of a land hemisphere and of a water hemisphere, as determined by Zenker, may, nevertheless, be accepted as the most trustworthy basis for further deductions.[1]

On the basis of the normal temperatures in a continental and in a marine climate, we may now try to determine the actual temperatures of the parallels of latitude in reference to their degree of "continentality," *i.e.*, of continental area. The temperatures of the parallels of latitude, which were thus determined by Zenker, showed, when compared with the values obtained by Dove and Spitaler from the isothermal charts, that the northern hemisphere beyond latitude 20° N. is warmer, while the southern hemisphere is somewhat cooler at all latitudes than should be the case, according to Zenker. This may be explained by the fact that a part of the heat received south of the equator is transferred to the northern hemisphere by ocean currents.

A comparison of the marine climates of the northern and southern hemispheres may properly be made at this point. This has been attempted by G. Neumayer, who compares the observations made on board of the *Challenger*, in the Pacific Ocean, in February, 1874, with those made on board of the *Vöringen* in the North Atlantic Ocean, in July, 1876-78. Both sets of observations relate to latitudes 60°-65°. In the case of latitude 55°, Neumayer uses his own observations in the South Pacific Ocean, in January, 1856, and compares them with the results of observations made on board of German vessels in the Atlantic Ocean, during July and August. The temperatures of air and water are, in both cases, 9°-10° higher in the North Atlantic than in the South Pacific, at the same latitude. The markedly low temperature of the South Pacific is clearly shown in these observations. As compared with the North Pacific Ocean, the difference would have been less, but still of the same character.

[1] On the basis of an assumption similar to that made by Zenker, and using the "thermal days" of Angot, Precht has calculated what he terms "ideal temperatures" for every 5° of latitude. Starting with the mean temperature of the earth as 15·1°, and the mean temperature of latitude 45°, 9·6°, Precht finds the transition from the "thermal days" to the mean temperatures. He computes the mean temperatures of the equator for the coefficients of transmission, 0·6, 0·67 and 0·7 as 26·4°, 26·5° and 26·7° respectively; those of the pole as −9·0°, −9·6° and −10·0°. These temperatures, as may be seen, would apply only in the case of a water hemisphere. (W. Precht: "Neue Normal Temperaturen," *M.Z.*, IX., 1894, 81-90.)

COMPARISON OF THE TEMPERATURE OF A MARINE CLIMATE IN THE NORTHERN AND SOUTHERN HEMISPHERES.

Air Temperature in Midsummer.

	Mean Latitude 64°.				Mean Latitude 55°.			
	Mean.	Diurnal Amplitude.	Maximum	Minimum.	Mean.	Diurnal Amplitude.	Maximum	Minimum.
	°	°	°	°	°	°	°	°
North, -	9·9	2·3	12·7	7·0	12·3	1·5	16·5	8·2
South, -	−0·6	2·5	2·8	−6·0	3·7	3·0	7·5	0·0

The ocean temperature, at latitude 64° N., was 10°, and at latitude 55° N., 9·0° higher than in corresponding latitudes of the southern hemisphere.[1]

Zenker's measure of "continentality."—In order to obtain a proper value for the derivation of temperature from the amount of radiation in a typical continental climate, Zenker was first obliged to establish the idea of "continentality" with reference to temperature conditions. For this purpose, the amount of annual temperature range serves, because this depends directly upon the difference of radiation in summer and in winter.

The annual range of temperature is not, however, itself any gauge of the degree of "continentality," because, as is well known, the former increases with latitude, even when the latter does not change. This increase in the annual range theoretically depends upon the sine of the latitude. The annual ranges of temperature must therefore be divided by the sine of the latitude in order to make them comparable. Zenker gives the reasons that made it seem to him more rational to divide by the arc instead of by the sine of the latitude.[2]

Zenker has calculated these ranges of temperature for all parts of the world, and has joined all places having equal ranges by lines. This gives a very clear picture of these ranges. The relative range of temperature is seen to be 100 per cent. at Verkhoyansk, north of Peking, and in the Sahara, north of Lake Chad. In the interior of Australia, the range reaches 80 per cent. only ; in the interior of North America it is 70 per cent. ; in Argentina, 40 per cent. ; in the interior of South Africa, 50 per cent. Over the oceans, 20 per cent., 10 per cent., and 0 per cent. are found. From the lines of equal relative ranges, by joining the values

[1] G. Neumayer : "Notwendigkeit und Durchführbarkeit der antarktischen Forschung," *Verhandl. V. Deutsch. Geographentags*, Hamburg, 1885, 172-196.

[2] If this is done, the relative value of the range of temperature at Vienna, for example, is 21·3° ÷ 48·2 = 0·44, or 44 per cent. Yakutsk, at latitude 62° N., with a January mean of −42·8°, and a July mean of 18·8°, has a relative range of temperature of 61·6° ÷ 62 = 99 per cent.

found for the continents and for the oceans, we obtain a gauge of the "continentality." The oceans, in their greatest extent in the northern hemisphere, show a relative range of 16 per cent., *i.e.*, about one-sixth of that in the interior of the continents.[1]

This conception of the degree of "continentality" of a climate, in reference to its temperature conditions, which is due to Zenker, is useful and important in climatology. Luigi di Marchi, in a recent publication,[2] likewise discusses the foregoing considerations. The derivation of the distribution of temperature over a land and a water hemisphere is carried out on purely theoretical grounds, and can therefore not be further considered here.

Distribution of pressure, rainfall, and cloudiness along parallels of latitude.—In connection with the foregoing extended discussion of the mean temperatures at different latitudes, it seems advisable to give also a general idea of the distribution of pressure, rainfall, and cloudiness. The mean annual pressures are those given by Ferrel,[3] and the monthly mean pressures are those recently determined by Baschin.[4]

It will be noticed that the means obtained from the January and July pressures do not give the annual mean, and it is not to be expected that they should, partly because of the character of the annual march of pressure, and, partly because the means given by Baschin are based on more recent observations. It is seen that there is a belt of low pressure between the equator and latitude 10° N.; two belts of high pressure in the sub-tropical latitudes between latitude 30° and latitude 40°, the belt in the southern hemisphere being somewhat nearer the equator than that in the northern; and then a decrease of pressure toward the pole, which is especially marked in the southern hemisphere.

[1] In order to determine the degree of "continentality," x, in percentages, when the relative range is n, we need to notice that $\frac{1}{6}(100-x)$ is the part which the marine climate plays in the annual range; and $1 \times x$ is the part played by the purely continental climate. Thus we have

$$x + \frac{1}{6}(100 - x) = n,$$

or

$$x = \frac{6}{5}n - 20.$$

The "continentality" of the climate of Vienna is, therefore, using the result already reached, $44 \times \frac{6}{5} - 20 = 33$; *i.e.*, one-third. The atmosphere in Vienna is likewise made up of one-third land air and two-thirds ocean air. The "continentality" of Yakutsk is $119 - 20 = 99$; that of Verkhoyansk is 100.

[2] *Le Cause dell' Era Glaciale*, Pavia, 1895, 100-124.

[3] W. Ferrel: *loc. cit.*

[4] O. Baschin: "Zur Frage des jahreszeitlichen Luftaustausches zwischen beiden Hemisphären," *Zeitschr. d. Gesell. f. Erdk.* (Berlin), XXX., 1895, 368-374.

DISTRIBUTION OF PRESSURE, RAINFALL, AND CLOUDINESS, BY PARALLELS OF LATITUDE.

Lat.	Mean Pressure at Sea Level. mm.			Annual Rainfall on Lands. cm.	Mean Cloudiness. Per Cent.
	Year. (Ferrel).	Jan. (Baschin).	July. (Baschin).		
° 80 N.	760·5	757·1	758·8	35	—
70	58·6	59·9	57·6	35	59
60	58·7	60·9	57·5	48	61
50	60·7	62·3	58·7	59	58
40	62·0	63·7	59·9	53*	49
30	61·7	64·9	59·0	60	42
20	59·2	62·7	57·9	82	40*
10	57·9	59·5	57·9	192	50
Equator.	58·0	58·0	59·4	195	58
10 S.	59·1	57·4	61·1	171	57
20	61·7	58·0	63·2	75	48
30	63·5	61·5	65·4	66*	46*
40	60·5	62·0	60·3	94	56
50	53·2	53·5	52·5	116	66
60	43·4	—	—	(102)	75
70	38·0	—	—	(89)	—

In the above table, the pressures are the actual pressures, reduced to standard gravity. The mean annual amounts of rainfall are those given by Sir John Murray.[1]

The amounts of rainfall concern the land areas of the world only, and therefore cannot be taken as representing the rainfall of the earth as a whole, especially not in the southern hemisphere. On the oceans, as is well known, measurements of rainfall have, until very recently, been almost wholly lacking, and, even now, our knowledge of rainfall over the oceans is in a very unsatisfactory condition. Important contributions to this subject have lately been made by Black and by Supan. Black[2] has discussed the question of measurements of rainfall made on

[1] John Murray: "On the total Annual Rainfall on the Land of the Globe and the Relation of Rainfall to the Annual Discharge of Rivers," *Scot. Geogr. Mag.*, III., 1887, 65-77. The English measures were converted into the metric system by Brückner. [*M.Z.*, IV., 1887, (63)-(64).] The values for the parallels of latitude were obtained from the values for the zones by graphic interpolation.

[2] W. G. Black: "Ocean Rainfall," *Journ. Manchester Geogr. Soc.*, XIV., 1898, 36-56 (New Edition, Edinburgh, 1899).

board ship, and has published a chart of ocean rainfall based on observations made on a number of British vessels equipped with marine rain-gauges, and Supan[1] has drawn isohyetal lines for the oceans, based on Black's data supplemented by observations taken on the vessels, *Novara*, *Gazelle*, and *Elizabeth*.

The amounts given for the continental interiors of the southern hemisphere are also very hypothetical. Nevertheless, there are certain general characteristics which may be looked upon as established, viz., the heaviest precipitation in the equatorial belt; the decrease toward the sub-tropical latitudes; the minima in the vicinity of the high pressure belts between latitudes 30° and 40°; an increase again as far as about latitude 50°; and then another decrease toward the pole. The last characteristic appears very distinctly in the northern hemisphere. The northern circumpolar region has a very small precipitation.

The mean cloudiness of the parallels of latitude is taken from the values given by Arrhenius, and by him obtained from de Bort's chart showing the distribution of cloudiness over the earth.[2] The minima of cloudiness in the sub-tropical latitudes appear very distinctly, as well as the large amount of cloudiness around the equator.

[1] A. Supan: "Die Vertheilung der Niederschläge auf den Meeren.," *Pet. Mitth.* XLIV., 1898, 179-182. Supan's isohyetal lines over the ocean are shown in Plate 18 of Bartholomew's *Atlas of Meteorology*.

[2] S. Arrhenius: "On the Influence of Carbonic Acid in the Air upon the Temperature of the Ground," *Philosoph. Mag.* (London), XLI., 1896, 275-276.

APPENDIX TO CHAPTER XII.

FORBES'S FORMULA FOR THE DISTRIBUTION OF TEMPERATURE OVER THE EARTH'S SURFACE.

The considerations which Forbes had in mind in constructing his formula for obtaining the mean temperatures of the parallels of latitude in the northern hemisphere may be briefly stated. The decrease of temperature with latitude along a meridian over the ocean is about in the ratio of the cosine of the latitude; while along a meridian over the land it is more rapid, being about as the square of this cosine. If, therefore, we choose an exponent, m, between 1 and 2, the mean temperature of a water hemisphere would be very well expressed by the formula

$$t_\phi = \text{temperature of the pole} + B \cos^m \phi.$$

As $\cos \phi = 1$ at the equator, the temperature at the equator = the temperature at the pole $+ B$. The influence of the land upon the mean temperature must be expressed by a third term, whose general form is to be determined by the fact that this term, as has already been seen, must become negative in the higher latitudes, because the presence of land there lowers the temperature, but must be positive in the lower latitudes. The change from one sign to the other is seen in the isothermal charts to take place between latitude 40° and 45°, where the temperature over the land becomes the same as that over the ocean. If we assume that the transition occurs just at latitude 45°, then the modifying influence of the land upon the temperature can be expressed as proportional to $\cos 2\phi$, because this expression becomes negative for a higher latitude than 45° ($2\phi > 90°$). If we denote by n the aliquot part of a parallel of latitude which is occupied by land, and by C a numerical coefficient which expresses the influence of the land upon the temperature in degrees of the thermometer scale, the temperature of a hemisphere which is covered partly by land and partly by water may be expressed by the formula

$$t_\phi = A + B \cos^m \phi + Cn \cos 2\phi.$$

The coefficients, A, B, C, and the exponent, m, must be derived from the observed mean temperatures of the parallels of latitude and from their

respective amounts of land area. By using the mean temperature obtained by Dove for latitudes 0°, 30°, 50°, and 70° N., Forbes finds

$$t_\phi = -10{\cdot}8° + 32{\cdot}9 \cos^{\frac{5}{4}}\phi + 21{\cdot}2n \cos 2\phi, \text{ in Centigrade degrees.}$$

This formula expresses the mean temperatures of the parallels of latitude as far north as latitude 75° perfectly satisfactorily. A further reason in favour of the general importance of this formula is the fact that it also accurately expresses the temperatures of the southern hemisphere as far as latitude 40° S., although it is based upon observations made in the northern hemisphere only, and the difference of the mean temperatures observed in corresponding latitudes of both hemispheres is considerable. For this reason, the formula may be regarded as being an approximate physical expression for the distribution of the air temperature at the earth's surface, and it therefore deserves mention in any text-book which deals with scientific climatology, even if the scope of the book is as limited as that of the present volume.

It will easily be seen that in the case of a water hemisphere, when $n=0$, the temperature of the pole, according to this formula, is $-10{\cdot}8°$, and that of the equator $32{\cdot}9° - 10{\cdot}8° = 22{\cdot}1°$. The former is, of course, to be regarded as a hypothetical temperature, which would be expected if there were open water on all sides as far as the pole, and if there were no large masses of ice present to radiate heat.

In the case of a land hemisphere (when $n=1$), the formula gives the temperature of the pole as $-10{\cdot}8° - 21{\cdot}2° = -32{\cdot}0°$, as $\cos^{\frac{5}{4}}\phi = 0$ and $\cos 2\phi = -1$. The temperature of the equator is found to be $-10{\cdot}8° + 32{\cdot}9° + 21{\cdot}2° = 43{\cdot}3°$. It will be seen that the coefficient of the third term is simply the difference between the temperature of the equator in a land hemisphere and on a water hemisphere. The mean temperature of both hemispheres, at latitude 45°, is found to be 10·6°.

By a second method, independent of the one just considered, Forbes determined the mean temperatures of the parallels of latitude on a land and a water hemisphere. The second method rests upon the fact, established by the first formula, that the influence of the land upon the mean temperature of a parallel of latitude may be considered proportional to the relative extent of that land. Forbes used the following equations in his second method: W is the temperature over the water; L is the temperature over the land; the coefficients of W and L are the relative proportions of land and water along the parallel.

Northern hemisphere.

$$\phi = 10°, \quad 26{\cdot}7 = 0{\cdot}748\,W + 0{\cdot}252L$$
$$\phi = 20°, \quad 25{\cdot}2 = 0{\cdot}682\,W' + 0{\cdot}318L'$$

etc.

Southern hemisphere.

$$\phi = 10°, \quad 25{\cdot}7 = 0{\cdot}790\,W + 0{\cdot}210L$$
$$\phi = 20°, \quad 23{\cdot}2 = 0{\cdot}795\,W' + 0{\cdot}205L'$$

etc.

By these equations, W and L may be determined for every latitude. The difficulty about this method of calculation is that the temperature of land and water on the same parallel must be determined on the basis of small differences.

For the temperatures of the parallels, as well as for the extent of land along the given parallels, the direct observed values are not employed, but smoothed values are used, which are obtained by graphic construction. This is a more rational method, because the temperature at latitude 10°, for example, is also influenced by the land masses somewhat north and somewhat south of that parallel, and not alone by the extent of land pertaining to that particular parallel.

B. MOUNTAIN CLIMATE.

CHAPTER XIII.

PRESSURE AND SOLAR RADIATION.

Mountain climate.—The second important subdivision of solar climate as modified by the special features of the earth's surface, is the climate which depends upon elevation above sea level. Mountain climates in all the zones have certain general characteristics which distinguish them from the climates of the surrounding lowlands. Elevation above sea level is, next to the distribution of land and water, the most important factor in causing differences of climate along the same parallel of latitude. Hence it is important to discuss briefly the meteorological phenomena which are observed everywhere as one ascends gradually from the lowlands on to mountains, and also to inquire into the general causes of these phenomena.

The decrease of pressure with altitude is the feature which shows the greatest regularity; so much so, that, as has already been pointed out, the pressure at any altitude can be calculated with a greater degree of accuracy, especially if the mean temperature of the air be known, than is really necessary in order that an opinion may be formed concerning the climate.[1]

[1] If h is the difference of altitude in meters; t, the mean temperature of the air column of the height, h; B, the known reading of the barometer at the lower level; b, the desired pressure at the upper station; then, with sufficient accuracy,

$$\log b = \log B - \frac{h}{18{,}460\,(1 + 0{\cdot}004\,t)}.$$

The denominator of the last term can be obtained from a table, as, *e.g.*, from Jelinek's *Anleitung zu meteorologischen Beobachtungen*, Part II., 1895, 47.

The following table gives the pressures corresponding to certain altitudes under the assumptions that the pressure at sea level is 762 mm., and that the decrease of temperature with altitude is uniformly at the rate of 0·5° in 100 m.

PRESSURES AT DIFFERENT ALTITUDES ABOVE SEA LEVEL.

Altitude (Meters).	Temperature at Sea Level.						Pressure change per 1°	Change of Altitude for each 1 mm. of pressure change.
	0°	5°	10°	15°	20°	25°		
	Mean Pressure in Millimeters.							
0	762	762	762	762	762	762	0·00	10·5
500	716	716	717	718	719	720	0·16	11·1
1000	671	673	675	676	678	679	0·32	11·8
1500	630	632	634	636	639	641	0·44	12·5
2000	590	593	596	599	601	604	0·56	13·4
2500	553	556	559	563	566	569	0·67	14·2
3000	517	521	525	529	532	536	0·76	15·1
3500	484	488	492	497	501	505	0·84	16·1
4000	452	457	461	466	470	475	0·91	17·2
5000	394	399	404	410	415	420	1·02	19·6
6000	343	348	353	359	364	370	1·09	22·5

It appears from the foregoing table that the mean pressure is not the same at the same altitudes in the tropics, in middle latitudes, and in higher latitudes. At an altitude of 3000 m., for example, the pressure is only 517 mm. at the temperature of 0°, but is 536 mm. at the temperature of 25°. Likewise, in the winter of central Europe, the pressure is about 517 mm. at 3000 m.; while in summer it is 532 mm. The column headed "Pressure change per 1°" shows in detail the influence upon the reading of the barometer of a change of 1° in the mean temperature of the air column. At an altitude of 3000 m. the effect of such a change in temperature amounts to nearly 0·8 mm. The last column on the right shows how far one must ascend in order to have the barometer fall 1 mm. While this distance amounts to about 10½ m. at sea level, it increases to 15 m. at an altitude of 3000 m.[1]

The following data as to the pressures at a few of the highest inhabited places in the world, give specific information concerning the

[1] Assuming a temperature of 0°. If the temperature is $t°$, then these figures must be multiplied by $1 + 0{\cdot}004\,t$, *i.e.*, they must be increased 0·4 % per degree, (or lowered if the temperatures are below freezing).

conditions of atmospheric pressure under which permanent human habitations are found. The mean annual temperatures at these stations are also given, although the subject of decrease of temperature with increase of altitude will later be considered in detail.

PRESSURES ON CERTAIN MOUNTAIN SUMMITS AND PLATEAUS.

Station.	Latitude.	Altitude. meters.	Pressure mm.	Temperature.
	° ′			°
Hospice of St. Bernard, - -	45 52 N.	2478	564	–1·3
Summit of Sonnblick, - - -	47 3 N.	3106	520	–6·3
City of Mexico, - - - -	19 25 N.	2270	586	16·3
Quito, - - - - - -	0 14 S.	2880	546	13·5
Leh (Tibet), - - - - -	34 10 N.	3517	497	5·7
Pike's Peak,[1] - - - -	38 50 N.	4308	451	–7·1
San Vincente.[2] - - - -	21 5 S.	4580	436	(3·0)
Hanle (Tibet), - - - -	32 48 N.	4610	433	(2·0)

On the interior plateau of Tibet there are said to be places over 4900 m. above sea level which are permanently inhabited; and in the Bolivian province of Chichas there is said to be a colony of miners living at an altitude of 5000 m.

Permanent human habitations are thus found at altitudes where the pressure is reduced to nearly one-half of the sea level pressure. According to Pöppig and Reck, the decrease in pressure at the highest inhabited places on the plateaus of the Andes has certain disagreeable physiological effects on the natives, as well as on persons who have but recently come to those great altitudes.

Physiological effects of diminished pressure.—The altitude above sea level at which mountain sickness (*Bergkrankheit*, *mal de montagne*, *puna*, *soróche*, *chuno*) is first noticed, varies greatly, because it depends upon the individual constitution, habits, attendant circumstances, bodily exertion, etc. In the Alps, the altitude at which most persons suffer from mountain sickness may be taken as between 3500 and 4000 m. Drew says that in Kashmir the people who live at an altitude of 1800 m. often suffer from mountain sickness after passing 3000 m. The Brothers Schlagintweit set the altitude of the beginning of mountain sickness in the Himalayas at 5000 m. This is also the greatest altitude at which sheep are pastured. Conway's expedition in

[1] Formerly occupied as a meteorological station.

[2] Near Portugalete, in Bolivia.

the Karakorum Mountains suffered from mountain sickness between 4900 and 5200 m., and Whymper and his companions on the slopes of Chimborazo, at 5100 m., which roughly corresponds to a pressure of 450-400 mm. On the other hand, Wolf and Whymper did not suffer at all on the summit of Cotopaxi, at 5960 m., although Whymper spent a night at that altitude, and was even free from mountain sickness at the summit of Chimborazo, to which he twice climbed (altitude 6250 m.; pressure, 358 mm.). Graham reached nearly 6700 m. in the Himalayas without having any difficulty in breathing; and similar exceptions not infrequently occur.[1]

In the establishment of the meteorological station of the Harvard College Observatory on the mountain, El Misti (5852 m.), in Peru, Bailey spent eight days consecutively at elevations between 3960 and 5790 m. During this time no real illness was experienced, although exercise was freely taken between 3960 and 4270 m., and at different times several busy hours were passed at 5790 m. without serious discomfort.[2] Recently, Conway has reached an altitude of 7393 m. on Mt. Sorata, in the Bolivian Andes, and reports that his party suffered more in the first struggles up to about 5500 m. than from there up to 6100 m. Above 6100 m. the difficulty again increased.[3] On Aconcagua, in Chile (7010 m.), Fitzgerald reports that he himself, and members of his party, suffered severely at various times at altitudes over 4900 m.[4]

Désiré Charnay notes that the Indians who bring down sulphur from the summit crater of Popocatepetl, and who therefore live at altitudes between 4000 and 5000 m., seemed strong and healthy, although they had been engaged in that occupation between 20 and 30 years. A similar statement is made by Griffith, in regard to the labourers on the Oroya railway, in Peru.[5]

[1] Schlagintweit: *Zeitschr. für Erdk.*, Berlin, 1866. Edward Whymper: *Travels amongst the Great Andes of Ecuador.* (A detailed account of mountain sickness on pp. 366-384, and in the appendix.) W. M. Conway: *Climbing and Exploration in the Karakorum-Himalayas*, London, 1894. *Scientific Reports*, C. S. Roy: *Mountain Sickness.*

[2] S. I. Bailey: "Peruvian Meteorology, 1880-1890," *Annals Astron. Obsy. Harv. Coll.*, XXXIX., Pt. I., Cambridge, 1899, 37. See also in connection with this, R. DeC. Ward: "A Visit to the Highest Meteorological Station in the World," *Monthly Weather Rev.*, April, 1898, 150-152; and "Sphygmograph Curves from 15,700 feet, and from 19,200 feet above Sea Level," *Journ. Boston Soc. Med. Sci.*, June, 1898.

[3] W. M. Conway: *The Bolivian Andes*, New York and London, 1901, 229-230, etc.

[4] E. A. Fitzgerald: *The Highest Andes*, New York, 1899, 55, 67, 77, etc.

[5] G. Griffith: "Mountain Sickness," *Nature*, LII., 1895, 414.

Pöppig and Reck describe, in some detail, the influence of the diminished pressure upon the human system in the high towns of the Peruvian and the Bolivian Andes. In Cerro de Pasco (4300 m.), every newcomer is at once attacked by mountain sickness. He has a feeling of suffocation; is subject to sleeplessness; and where the street is somewhat steep, pulls himself along the houses with difficulty, stopping to rest at corners and in doorways. The night is the time of the worst suffering, fainting spells occurring now and then. At the end of six or seven days, everyone with sound heart and lungs recovers; but the after-effects may be noticed for some weeks. Under the influence of the wind, the skin cracks open; blood runs from the lips and nose; and at night, face and hands swell. If often experienced, the *chuno* leaves black furrows on the fingers, which serve as a ready means of identifying the inhabitants of the higher altitudes in the Andes, just as the Indians of the forests are recognised by the black dots on their skin, caused by mosquito bites.[1]

That mountain sickness is much increased by wind is abundantly confirmed by the brothers Schlagintweit.

There is abundant evidence that mountain sickness is not merely a result of the exertion necessary in climbing. Even when altitudes of about 4000 m. are reached on horseback or by train, mountain sickness is experienced. Ball, a skilled mountain climber, suffered from it at Chicla, the former terminus of the Oroya railroad from Lima, at an altitude of 3720 m.; and many of those who visit Pike's Peak (4308 m.) by means of the cog-wheel railway, also complain of mountain sickness.

Abercromby, who rode on horseback to the top of Pike's Peak, reports that he was not affected immediately after his arrival, but began to suffer during the night and on the following morning, while those who walked up experienced mountain sickness immediately. The observers at the meteorological station on Pike's Peak likewise suffered from mountain sickness, and had frequently to be relieved of duty there. Some of the men could not be acclimated at all, and had to leave the mountain immediately.

According to Dr. Egli-Sinclair, Jansen suffered also, although he was carried to the summit of Mont Blanc; as did Dr. Kronecker and his companions on the Breithorn, at 3750 m., under similar circumstances.[2]

Bailey reports that in going from Arequipa to Crucero Alto, in southern Peru (2301 m. to 4470 m.) many persons suffer slightly, and

[1] Pöppig: *Reisen*; I. H. Reck: "Geographie und Statistik der Republik Bolivia," *Pet. Mitth.*, XII., 1867, 243-251.

[2] R. Abercromby: *Seas and Skies in many Latitudes*, London, 1888, 406. J. Ball: *Notes of a Naturalist*, London, 1887, 81; "Notes on Soroche in the Andes," *Nature*, XXVI., 1882, 477-478.

some quite seriously, especially if they remain over night at the latter station.[1]

Rotch[2] reached an altitude of 5075 m. on Mount Charchani, in Peru, without any physical exertion, the ascent being made on mule-back; and although no nausea or severe headache was experienced, other symptoms manifested themselves in abnormal excitability and restlessness, making sleep impossible, and in a lapse of memory and a want of sequence of ideas.

Ward[3] went to the summit of El Misti (5852 m.) on mule-back, but on arrival the feeling of complete exhaustion was so great that for an hour and a half the slightest exertion was out of the question. There was some tendency to faintness during this time, and nausea and headache were also experienced. Assistance was necessary in remounting the mule before making the descent. On a second visit to the summit of El Misti, the same writer reports feeling fairly well when lying down, but the exertion of walking even a few steps brought on a feeling of exhaustion and nausea, and increased an already severe headache. Upton,[4] on the summit of El Misti, noted a *decrease* in his respiration.

Altitudes reached by balloons.—The altitudes reached by mountain climbers have not been greater than about 7400 meters. Thus, Schlagintweit reached 6780 m., and a pressure of 339 mm. on the peak of Ibi Gamin; and Conway came very near 7400 meters on Mount Sorata. By means of balloons, however, heights of 8000 to 10,000 m. have been reached. On September 5, 1864, Glaisher reached a height of about 8600 m., with a pressure of 248 mm., when he became unconscious. Berson, on December 4, 1894, reached 9100 m., where he found a pressure of 231 mm., and a temperature of –48°. He was able to keep himself well and active by breathing oxygen, the need of which began to be felt at 6700 m. On August 1, 1901, Berson and Süring reached 10,300 m., and recorded a temperature of –40°. The tragic fate of the three passengers in the balloon, "Zenith," is well known. Two of these, Crocé-Spinelli and Sivel, died on April 15, 1875, at an altitude of about 8500 m., when the

[1] *Peruvian Meteorology*, 36, 37.

[2] A. L. Rotch: "The Highest Meteorological Station in the World," *Am. Met. Journ.*, X., 1893-4, 286.

[3] R. DeC. Ward: "A Visit to the Highest Meteorological Station in the World," *Boston Med. and Surg. Journ.*, CXXXVII., 1897, 637-39.

[4] W. Upton: "Physiological Effects of Diminished Air Pressure," *Science*, N.S., XIV., 1901, 1012-1013.

pressure was 260 mm., and the amount of oxygen in the air which they breathed was reduced to one-third of the amount at sea level. Tissandier, the third occupant of the car on this memorable ascent, very nearly met the same fate as that of his companions, although a supply of oxygen had been taken up. The time spent at such great altitudes in the car of a balloon is, however, too short to furnish data for a study of the phenomena and causes of mountain sickness.

Symptoms of mountain sickness.—The chief, as well as the most common, symptoms of mountain sickness are the following:—A craving for air; dizziness; increasing shortness of breath; weakness of the muscles; diminished powers of endurance; lack of energy; indifference to surroundings and to danger; nose-bleed; palpitation of the heart; headache; occasional nausea; loss of appetite. Respiration is quick and irregular, and if the symptoms become greatly exaggerated, unconsciousness, and even death, follow.

Causes and consequences of mountain sickness.—Of the earlier investigations into the effects of diminished pressure upon the animal organism, those of Jourdanet are especially noteworthy. Jourdanet found the inhabitants of the high plateau of Anahuac by no means so strong and so energetic as he expected to find them, in view of the low temperature on the plateau as compared with that on the Mexican lowlands. The inhabitants of this plateau have a calm, resigned, thoughtful temperament; their complexions are yellow or pale; their muscles are flabby, and they are susceptible to disease. All the physiological appearances point to an anaemic condition. Jourdanet concluded that these symptoms were due to a decrease in the oxygen of the blood, and this condition he terms *anoxyhemia*.[1]

The experiments of Paul Bert confirm these conclusions.[2] These experiments showed that the effect of diminished pressure remains imperceptible until the pressure of the oxygen has decreased by one-fourth, *i.e.*, until the air pressure of 760 mm. has decreased 190 mm., or has reached about 570 mm. This takes place at about 2000 m. above sea-level. At this altitude the influence of the decrease in oxygen pressure becomes noticeable through the decreased oxygenation of the blood, with all the consequences that result therefrom. The cause of the weakness of the inhabitants of great altitudes is therefore to be sought in the insufficient oxygenation of the blood in a rarefied

[1] Jourdanet: "Influence de la Pression de l'Air sur la Vie de l'Homme; Climats d'Altitude et Climats de Montagne," Paris, 1876.

[2] Paul Bert: *La Pression Barométrique*, 1878.

atmosphere. Whymper also shows that there was a decrease in the physical powers of himself and his companions on the high plateau of Ecuador.[1] This effect not being noticeable below 2000 m., Jourdanet suggests that the climates of mountainous regions may be divided into *climats de montagne* below 2000 m., and *climats d'altitude* above that height.

During a stay of several days on Mont Blanc, in the Vallot Observatory (4400 m.), Dr. Egli-Sinclair was able to prove, in his own case and in that of his companions, the decrease in the amount of haemoglobin, or red colouring matter in the blood, which had been suspected by Jourdanet. All the party, including the veteran mountain climber Infeld, suffered from mountain sickness during the first three days, the third day being the worst. After that, the condition of the party began to improve, and the amount of haemoglobin in the blood of all the members increased again.[2]

The experience of Rotch during three visits of from one to three days' duration at M. Vallot's cabin, where Dr. Egli-Sinclair's observations were made, corroborates the latter. The sickness did not appear at once, but after several hours' rest. It was more severe at night, and when reclining, and it was alleviated by the inhalation of oxygen and by taking phenacetine. The acute symptoms disappeared on the second day, although respiration remained difficult.[3]

Viault examined the blood of acclimated men and of animals on the Pic du Midi (2877 m.), and on the plateaus of Peru, at altitudes of 3700 to 4400 m. As a result of these studies, it appeared that there was a very much greater amount of haemoglobin in the blood than usual, a fact which enables the blood to absorb sufficient oxygen from the rarefied air. Viault's own blood contained five millions of red blood corpuscles per cubic millimeter while he was in Lima. After a stay of fourteen days at Maracocha, at an altitude of 4400 m., the number rose to seven millions, and a week later, to eight millions. The examination of the blood of Viault's companions gave similar results. The increase in the numbers of blood corpuscles in the case of animals which were taken from the lowlands to the plateau, was in the ratio of 4·8 to 7·0. Müntz found almost the same ratio in the case of

[1] *Loc. cit.*, 373. Compare also the interesting discussion by R. M. Bosanquet on "Mountain Sickness, and Power and Endurance," *Philosoph. Mag.*, London, XXXVIII., 1893, 47-52.

[2] *Jahrb. d. Schweiz. Alpen Club*, XXVII., 308.

[3] A. L. Rotch: "Physiological Effects of High Altitudes," *Am. Met. Journ.*, XII., 1895-96, 222.

rabbits which were taken on to the Pic du Midi, even after they had been there six weeks. The amount of iron in the blood rose in the ratio of 4 : 7. When men and animals descend to the lowlands again, they lose this increased haemoglobin in their blood. Mountain sickness is, therefore, only the first phase of the battle of the organism with the changed conditions of life, namely, the rare atmosphere. To these changed conditions the organism is soon able to adjust itself.[1]

Roy also comes to the conclusion that all the symptoms of mountain sickness can be attributed to asphyxia, *i.e.*, to the decrease in the amount of oxygen which is furnished to the tissues. The same result is noted at lower altitudes, where the supply of oxygen has not been so greatly decreased, provided there is increased need of oxygen in the tissues, as occurs when there is muscular exertion.[2]

The fact that snow increases the tendency to mountain sickness has been frequently observed. It seems that this sickness is less likely to occur where there is no snow on the ground. Roy attributes this to the fact that snow, when wet, absorbs much oxygen.

Pressure changes on mountains.—The irregular changes in pressure decrease with altitude in the same ratio as the pressure itself decreases; but hardly any importance attaches to this fact from a climatic point of view. The mean monthly ranges of pressure on the summit of the Schafberg, for example, at an altitude of 1780 m., or about that of the Upper Engadine, amount in winter to 19 mm.; while at the base, at Ischl (467 m.), the range is 25·0 mm. In summer, the ranges are 12·0 mm. at the summit, and 14·4 mm. at the base. The annual changes in pressure at greater altitudes are characterised by the fact that the pressure is lowest in winter and highest in summer. This results from the changes in the temperature of the atmospheric strata, as is shown in the table on page 223. The annual range of pressure, therefore,

[1] F. Viault: "Sur l'Augmentation considérable du Nombre des Globules rouges dans le Sang chez les Habitants des hauts Plateaux de l'Amérique du Sud," *Comptes Rendus*, CXI., 1890, 917-918; "Sur la Quantité d'Oxygéne contenue dans le Sang des Animaux des hauts Plateaux de l'Amérique du Sud," *ibid.*, CXII., 1891, 295-298; A. Müntz: "De l'Enrichessement du Sang en Hémoglobine suivant les Conditions d'Existence," *ibid.*, 298-301; F. Viault: "Action physiologique des Climats de Montagne," *ibid.*, CXIV., 1892, 152-155.

[2] W. M. Conway, *loc. cit.* See also recent publications by Paul Regnard: "La Cure d'Altitude," 8vo., Paris, 1897, pp. 435 [*M.Z.*, XV., 1898 (42)-(43)], and A. Mosso: "Der Mensch auf den Hochalpen," 8vo., Leipzig, 1899, pp. 483 (*M.Z.*, XVI., 1899, 282-285).

also increases with increase of altitude, as is seen in the following examples:—

ANNUAL RANGES OF PRESSURE AT DIFFERENT ALTITUDES (MM.).

Station.	Altitude (m.)	Minimum.	Maximum.	Difference.
Geneva, - -	405	724·7 (March).	728·0 (Jan. & Sept.)	3·3
St. Bernard, -	2476	559·3 (,,).	768·5 (July).	9·2
Sonnblick, - -	3100	514·4 (,,).	525·0 (,,).	10·6
Colorado Springs,[1]	1856	607·1 (,,).	613·7 (August).	6·6
Pike's Peak,[1] -	4308	443·4 (Feb.).	459·6 (July).	16·2

The diurnal range of pressure in valleys in mountainous regions agrees, in general, with that over lowlands, but the afternoon minimum is very pronounced. The southern Alpine valleys have a diurnal range which is almost tropical in its character, the difference between the barometer readings at 7 or 8 o'clock A.M., and in the afternoon, being 2 mm. and more. On mountain summits and slopes the morning minimum becomes the chief minimum of the day, while the afternoon minimum is much less marked, and comes later, than on the lowlands.

Increase in intensity of insolation with increase of altitude.—As the atmospheric strata which absorb the radiant energy from the sun become less dense with increasing elevation above sea-level, it follows that there must also be less absorption aloft. In other words, the intensity of solar radiation must increase with increasing altitude. Thus at Leh, for example, the thickness of the absorbing atmospheric stratum has decreased by one-third, as may be seen in the table of pressures given on page 224. Furthermore, water vapour is a better absorber of solar radiation than dry air, and decreases with altitude more rapidly than the pressure decreases. Hence it follows that the intensity of the total solar radiation increases more rapidly with increasing altitude than would be inferred from the rate of decrease of pressure alone. This rapid increase of solar radiation must be especially prominent near the upper limit of the aqueous atmosphere. A large part of the band absorption of solar rays is exercised in an elevated atmospheric zone above an altitude of 10,000 m. (perhaps 10-30 km.), where, as de Bort has found, the heat produced by the sun's

[1] Corresponding mean pressures.

rays nearly obliterates the normal altitudinal temperature gradient. On August 11, 1867, Cayley saw a thermometer, exposed in bright sunshine at Leh, rise to 57·8°, whereas the temperature in the shade was only 23·9°. A black-bulb thermometer *in vacuo* even rose to 101·7°, *i.e.*, nearly 14° above the boiling-point of water, which at the altitude of Leh is only 88°. At altitudes between 3000 and 4600 m. in the Himalayas, Hooker saw the black-bulb thermometer in the sunshine rise 40°-50° above the shade temperature in winter, which is the dry, clear season in that region. On one occasion the black-bulb thermometer stood at 55·5° at 9 A.M., while the temperature of a shaded snow-surface at the same time was – 5·6°. The great intensity of the sun's rays attracts the attention of everyone who goes to considerable altitudes. Saussure first directly proved this increased intensity, and he was followed by a considerable number of other investigators, namely, Bravais and Martins, Forbes, Soret, and more recently by Violle, Langley, and others.

Temperatures in sunshine and in shade at different altitudes: Frankland.—The results obtained by absolute measurements may well be preceded by some recent measurements made by Frankland in regard to the increase with altitude of the difference between the temperature in the shade and in the sunshine. The temperature in the sunshine is that obtained by a black-bulb thermometer *in vacuo.* The sun's altitude was constant at 60°.

Station.	Altitude (m.).	Thermometer in		Difference.
		Shade.	Sun.	
		°	°	°
Oatland Park, - -	46	30·0	41·5	11·5
Riffelberg, - - -	2570	24·5	45·5	21·0
Hörnli, - - - -	2890	20·1	48·1	28·0
Gornergrat, - - -	3140	14·2	47·0	32·8
Whitby, - - - -	20	32·2	37·8	5·6
Pontresina, - - -	1800	26·5	44·0	17·5
Bernina Hotel, - -	2330	19·1	46·4	27·3
Diavolezza, - - -	2980	6·0	59·5	53·5

Violle's results for Mont Blanc and the Bossons Glacier.—On August 16 and 17, 1875, Violle made a series of absolute measurements of solar radiation on the summit of Mont Blanc, and at the Grands Mulets, while Margottet simultaneously made similar measure-

ments with the same sort of apparatus at the foot of the Bossons Glacier. The results were as follows :—

	Altitude. meters.	Pressure. mm.	Vapour Tension. mm.	Intensity of Solar Radiation.[1]
Mont Blanc, - - -	4810	430	1·0	2·39
Grands Mulets, - -	3050	533	4·0	2·26
Bossons Glacier, - -	1200	661	5·3	2·02

The intensity of solar radiation on the summit of Mont Blanc was thus 15 per cent. greater than at the Bossons Glacier, 3600 m. lower; and 26 per cent. greater than at the level of Paris (60 m.). These figures are all reduced to a zenithal position of the sun, or a thickness of the atmosphere = 1.

Violle concludes from his observations that five times as much radiant energy is absorbed as heat by the water vapour of the atmosphere as by the dry atmosphere, and that, therefore, in view of the fact that the quantity of water vapour in summer amounts to but about 1/380th of the mass of the atmosphere, the coefficient of absorption of water vapour is 1900 times greater than that of air.[2]

Effect of water vapour on atmospheric absorption of solar radiation.—The water vapour of the atmosphere has a very marked influence upon absorption, especially in the case of the less refracted rays from yellow to red, and beyond. This has already been proved by many physicists and meteorologists, partly by direct observation and partly as an inference based upon the corresponding behaviour of water toward these same kinds of rays. Langley, Abney, and others have shown that there are strong absorption bands in the infra-red portion of the spectrum,[3] as has already been pointed out on page 121. Even the carbon dioxide of the atmosphere, notwithstanding its small quantity, has some effect upon atmospheric absorption of solar radiation.

The rapid increase in the intensity of solar radiation with increase of altitude in the lower part of the atmosphere is due not so much to the diminution of water vapour as we ascend, for the infra-red vapour

[1] The unit in the last column of the table is the value of the radiant energy received on a surface of one square centimeter in one minute.

[2] See also J. M. Pernter : "Absolute Messungen der Sonnenwärme auf dem Monte Rosa und auf dem Mont Blanc," *M.Z.*, XV., 1898, 105-108.

[3] See recent investigations of Langley and Very, already referred to, and also W. C. Röntgen : "Neue Versuche ueber die Absorption von Wärme durch Wasserdampf," *Wied. Annalen*, XXIII., 1884, 1-49 ; 259-298 ; also F. Paschen : "Ueber die Emission erhitzter Gase," *ibid.*, L., 1893, 409-443 ; LI., 1894, 1-39.

bands in the spectrum still continue strong, but is largely attributable to the cessation of atmospheric dust (including under this term aqueous condensation products), which affects chiefly the shorter waves, whence these are especially strong on mountain tops. The loss of heat by radiation increases in proportion to the decrease of water vapour, because the strongest terrestrial radiation falls within the limits of the principal water bands. Terrestrial radiation, therefore, increases at high altitudes.

In a preliminary report upon the results of the Mt. Whitney expedition (August, 1881), which was undertaken for the purpose of studying the absorption of solar radiation by the earth's atmosphere, Langley mentions the extraordinary dryness and transparency of the air at an altitude of 4000 to 4500 m. during the Californian summer. It was impossible to gauge distances with any degree of accuracy, owing to the absence of atmospheric perspective. The intensity of solar radiation on the summit of the mountain was so great that the temperature in a copper vessel, covered with two glass plates, rose far above the boiling-point. Face and hands were burned by the intense insolation, although the surface of the mountain was not snow or ice, but rock. The sky was perfectly clear, and its colour was a deeper violet than Langley had ever noticed, even on the top of Mt. Etna. The transparency of the air on the Tibetan plateaus likewise surprised Schuster. It was impossible to judge of distance; the eye seemed to be looking through a vacuum. The characteristic bluish colour, which gives a charm to the distant landscape near sea-level, was almost altogether lacking. There was a remarkable clearness in the blue of the sky. Schuster was especially surprised at finding hardly any reddening of the clouds at sunset. When the sunset colours did appear they were yellow rather than red. In Simla, however, as he learned, the red sunset colours were often and beautifully seen at the end of the rainy season.[1] It is therefore the same deficiency of water vapour, and of the products of its condensation, that brings about these phenomena.

A marked increase in the intensity of the ultra-violet rays at great altitudes has been proved by Langley, Abney, and Simony (on the Peak of Teneriffe), as well as by Elster and Geitel. The decrease in the intensity of the ultra-violet rays is most marked in the lower strata of the atmosphere, up to 1500 to 2000 m., while at greater altitudes there is less decrease. Elster and Geitel found the following ratios for the increase in the ultra-violet rays (J) with increasing altitude :—

Station, - -	Wolfenbüttel,	Kolm Saigurn.	Sonnblick.	Upper Limit of Atmosphere.
Altitude, -	80 meters.	1600 meters.	3100 meters.	—
J, - - -	38	72	94	236

[1] A. Schuster: "Scientific Notes taken in the Himalayas," *Nature*, XIII., 1875-76, 393-395.

Between the upper limit of the atmosphere and the top of the Sonnblick at 3100 m., 60 per cent. of the ultra-violet rays are lost. Between the top of the Sonnblick and Kolm Saigurn, at 1600 m., 23 per cent. of the remainder are lost, and in the lower 1600 m., 47 per cent. of the last value are lost.[1]

The great intensity of solar radiation, and especially of the ultra-violet rays, has a considerable influence upon vegetation at great altitudes. The fact that the skin is much tanned and burned at great altitudes is probably due to the same cause, although reflection from the snow, as well as the dryness of the air, play a part in this process. Persons who spend the winter at Davos Platz have their faces well burned by the sun. Tyndall notes the fact that his skin was never burned more than when he was working in a lighthouse under an electric light, the latter being, as is well known, very rich in blue and ultra-violet rays.[2]

The increase in the chemical effects of sunlight with increasing altitude was thoroughly studied by Bunsen and Roscoe a number of years ago, and the chief results obtained by them are given in the table below. In this table, the intensity of the chemical effects of sunlight is expressed in percentages of the intensity just outside of the earth's atmosphere. Bunsen and Roscoe computed the latter to be equal to 35·3 light units, *i.e.*, the radiation from the sun would here be strong enough, if falling at normal incidence on an infinite column of mixed hydrogen and chlorine, to create a layer of hydrochloric acid gas of so many meters in thickness in one minute.

CHEMICAL INTENSITY OF SUNLIGHT
(IN PERCENTAGES OF THE MAXIMUM).

Pressure. mm.	Altitude above Sea-level. m.	Sun's Altitude.				
		90°	70°	50°	30°	10°
750	130	44	42	34	19	1
650	1270	49	47	39	24	2
550	2600	55	53	46	30	3
450	4200	61	59	53	37	6
350	6200	68	67	61	46	10

[1] J. Elster and H. Geitel: "Beobachtungen betreffend die Absorption des ultravioletten Sonnenlichtes in der Atmosphäre, *M.Z.*, X., 1893, 41-49.

[2] See Dr. Robert L. Bowles: *Sunburn on the Alps*, London, 1890. *Mitth. des Deutsch. u. Oesterreich. Alpenver.*, 1890, 78.

It is seen from the above table that at an altitude of 2600 m. the chemical power of the solar radiation which corresponds to the more refrangible portion of the spectrum is 11 per cent. greater than at sea-level, when the sun is in the zenith, and is two or three times as great as at sea-level, when the sun is low.

"At the time when the sun has nearly reached the zenith in the latitude of the Himalayas, the amount of the direct sunlight which falls on the valleys of the Thibetian highlands, where grain is cultivated, is nearly one and a half times as large as that falling on the neighbouring lowlands of Hindostan. This difference increases in so rapid a ratio with increasing zenith-distance, that when the sun is 45° removed from the zenith, the direct solar rays on the high table-land of Thibet give more than twice the chemical action of those falling on the plains of India."[1]

Surface temperatures on mountains.—Closely associated with the great intensity of solar radiation, which has been shown to be an important characteristic of mountain climates, is a relatively high surface temperature. With increasing altitude above sea-level, the excess of surface temperature over air temperature is likewise found to increase. In making any estimate of the effects of mountain climates upon plant life, some account must be taken of the surface temperature, as well as of the temperature of the air. The importance of the surface temperatures may be gauged by means of the observations which follow. According to von Kerner's observations in the Central Alps of the Tyrol,[2] the mean difference between air temperature and surface temperature at 1000 m. is 1·5°; at 1300 m., 1·7°; at 1600 m., 2·4°. In Alsace, Boller finds this difference 1·0° at Hagenau, at an altitude of 145 m., and 1·3° at Melkerei, altitude 930 m. In summer, these differences rise to 2° or 3°. Martins gives the following comparison of his observations of the temperature of the air and of the surface on the Faulhorn and at Brussels.

TEMPERATURE ON THE FAULHORN AND AT BRUSSELS AT 9 A.M., AUGUST 10–18, 1842.

Station.	Altitude m.	Air.	Surface of Ground.
Faulhorn, -	2680	8·2°	16·2°
Brussels, -	50	21·4	20·1

[1] R. Bunsen and H. E. Roscoe: "Photochemical Researches," Part IV., *Philosoph. Trans. Roy. Soc.*, London, CXLIX., 1859 (879-926), 914.

[2] A. Kerner: *The Natural History of Plants* (trans. by F. W. Oliver), Vol. I., London, 1894, 525.

The surface temperature on the Faulhorn was thus only 4° lower than that at Brussels, while the air temperature shows a difference of more than 13°. The mean shade temperature of the air on the Faulhorn was 6·7°; that of the surface of the ground, 9·5°; and that at a depth of 1 dm. under ground, 10°. The mean maximum air temperature was 9·0°; that of the surface of the ground was 19·5°.

According to observations made from September 21 to October 4, 1844, the mean temperature of the air was 5·4°; while that of the surface was 11·8°. The summit of the Faulhorn is just a little below the snow-line, as is the case also with Magdalen Bay, on the island of Spitzbergen (80° 34′ N.). While the surface temperature is much higher than the air temperature in the case of the Faulhorn, the surface temperature in the Spitzbergen case is 1° below the air temperature. The explanation of this fact is to be found in the intense solar radiation on mountain tops.

In 1864, Martins carried out a series of simultaneous observations of air and surface temperatures, on the top of the Pic du Midi (2877 m.) and at Bagnères (551 m.). The observations were made on three perfectly clear days, September 8, 9, and 10; and the lower station was but 14·5 km. in a direct line from the higher. The observation of the surface temperature was carried out in precisely the same manner at both stations, and the same soil was chosen in both cases, namely, black loam composed of the decayed trunks of old willows. The results are as follows:—

AIR TEMPERATURES AND SURFACE TEMPERATURES AT BAGNÈRES AND ON THE PIC DU MIDI.

	Bagnères.	Pic du Midi.	Difference.
	°	°	°
Mean Air Temperature,	22·3	10·1	12·2
Mean Surface Temperature.	36·1	33·8	2·3

The temperature of the ground at a depth of 5 cm. was 25·5°, at Bagnères, which was 3·2° higher than the air temperature; while on the Pic du Midi, the former temperature was 17·1° or 7·0° higher than the latter. Thus the earth's surface on the Pic du Midi, down to a depth of several centimeters, was heated about twice as much as at Bagnères, 2326 m. nearer sea level. The absolute maxima were as follows:—Bagnères, September 9, 2 P.M., surface, 50·3°; air, 27·1°. Pic du Midi, September 10, 11.30 A.M., surface, 52·3°; air, 13·2°. As the summit of the Pic du Midi always became covered with clouds toward

noon, the maximum surface temperature was reached before noon, and was 2° higher than at Bagnères.

The high surface temperature and the great intensity of the sunshine are favourable characteristics of mountain climates, as compared with the climates of polar regions which have the same air temperature. Thus it comes about that the summit of the Faulhorn, with a surface of $4\frac{1}{2}$ hectares, has a total of 131 species of phaenogamous plants, while the whole archipelago of Spitzbergen is said to have only 93 species. The long day cannot make up for the lower intensity of solar radiation; the surface temperature is no higher than the air temperature, and the ground remains frozen at a depth of a few decimeters.

Nocturnal radiation on mountains.—The rarity of the air, and the decrease in the amount of water vapour with increasing altitude, involve a more active radiation of heat at night, as well as more intense insolation by day. Comparative measurements of radiation at Brienz and on the top of the Faulhorn, 2100 m. higher, made with Pouillet's actinometer, showed 37 per cent. greater radiation at the last-named station. Similar observations, made simultaneously at Chamonix and on the Grand Plateau of Mont Blanc (3930 m.), showed that the radiation was almost twice as great (93 per cent. greater) at the latter station, which is 2880 m. higher. The temperature of the snow on the Grand Plateau fell during the nights of August 28-31, 1844, to – 19·2°, while the air temperature was only – 6·5°, according to Martins. In February, Pernter found the nocturnal radiation on the Sonnblick (3100 m.), 0·20 calories, while Maurer found it to be only 0·13 calories at Zürich (450 m.). On the Pic du Midi, also, the temperature minima on the surface (September 8 and 9), were 13·1° lower than those at Bagnères; while the maxima on the mountain equalled, or exceeded, the maxima at the lower station. As the gain and loss of energy by absorption and radiation are both increased on mountains, it follows that there is a much greater range of surface temperature from day to night on mountains than on lowlands.

Importance of exposure in controlling surface temperature in mountain climates.—In all mountain climates the exposure has a marked influence on the insolation, and therefore upon the surface temperatures. With the exception of the region about the poles, where the sun makes a complete circuit of the horizon, there are, in each hemisphere, slopes which are at a peculiar advantage, or at a peculiar disadvantage, as regards the amount of insolation which they receive. In the northern hemisphere, because of the higher angle at which the sun's rays strike them, the southern slopes receive the stronger insolation; while in the

southern hemisphere, the northern slopes are the most favourably situated. Furthermore, when the sun's altitude is low, the land which slopes away from the sun is in the shade a longer time, and this fact still further increases the excess of insolation received by southern slopes in the northern hemisphere. Along the equator, northern and southern slopes are equally favoured; while on the other hand, eastern and western slopes are in the shade somewhat longer. Valleys which trend east and west are more favourably situated than those which trend north and south, provided the angle of slope of the enclosing mountain sides is the same, because the north and south valleys are in the shade longer than the east and west valleys. At noon, the sun is high enough to be above the mountain tops almost everywhere, but the east and west valleys also have the morning and the evening sun. In mountainous regions, marked climatic contrasts within short distances result from differences in situation as regards shady or sunny slopes. The southern slopes may be decked with a mantle of green in the early spring, while the northern slopes are still covered with snow. In summer, the sunny slopes may be gorgeous with yellow fields of ripening grain, while the northern slopes are covered with dark forests of evergreen.

In addition to the differences in insolation on the different slopes of mountains, there is also the influence of winds of varying temperatures. The effect of the winds is, in general, similar to that of the exposure, *i.e.*, to warm the southern and to cool the northern slopes in the northern hemisphere. Unfortunately, there are hardly any observations of the effect of exposure, which is an important climatic factor. If such observations were available, numerical data could be given regarding the difference in surface and air temperatures as dependent upon the direction of the slope of mountain sides in different latitudes.

Soil temperatures under different exposures.—We have, however, a valuable series of observations of soil temperatures, made by A. von Kerner under different conditions of exposure as regards direction of slope. These observations were carried on during three years, at a depth of 80 cm., on an isolated hill near Innsbruck, at about 600 m. above sea-level. Later, similar observations were also made for three years at Trins, in the Gschnitz valley (1340 m.), and these have been discussed by F. von Kerner.[1]

[1] A. Kerner: "Ueber Wanderungen des Maximums der Bodentemperatur," *Z.f.M.*, VI., 1871, 65-71. F. von Kerner: "Die Aenderung der Bodentemperatur mit der Exposition," *S.W.A.*, C., II a, 1891, 704-729. E. Wollny: "Untersuch-

It appeared, in substantial agreement at both altitudes, that the surface with the southwestern exposure has the highest temperature in winter, while that with a southeastern and southern exposure has the highest temperature in summer. The mean soil temperatures were as follows:—

SOIL TEMPERATURES AT A DEPTH OF 80 CM. ON SLOPES WITH DIFFERENT EXPOSURES.

Slope.	Inn Valley (600 m.).			Gschnitz Valley (1340 m.).		
	Winter.	Summer.	Year.	Winter.	Summer.	Year.
	°	°	°	°	°	°
N., - -	4·2	15·3*	9·5*	0·6	11 2*	5·1*
N.E., -	4·4	17·0	10·6	0·9	11·6	5·5
E., - -	4·0*	18·6	11·3	0·4*	12·6	5·9
S.E., -	5·1	**19·7**	12·6	1·5	13·4	7·5
S., - -	5·3	19·3	**12·6**	2·4	**13·4**	**7·8**
S.W., -	**6·6**	18·3	12·7	**3·1**	12·9	7·8
W., - -	5·5	18·5	12·2	2·6	12·6	7·4
N.W., -	4·5	16·0	10·2	2·0	11·9	6·5
Mean, -	5·0	17·8	11·5	1·7	12·5	6·7
Range, -	2·6	4·4	3·2	2·7	2·2	2·7

The migration of the maximum soil temperature, from S.W. in winter to S.E. in summer, is probably due chiefly to the diurnal variation of cloudiness. In winter, the sun dissipates the clouds, which are chiefly

ungen ueber den Einfluss der Exposition auf die Erwärmung des Bodens," *Forsch. auf dem Geb. der Agrikulturphysik*, I., 1878, 263; and "Untersuchungen ueber die Feuchtigkeit und Temperatur des Bodens bei verschiedener Neigung und Exposition desselben," *ibid.*, X., 1887, 1 and 345. In the latter treatise, Wollny also discusses the daily march of temperature under different conditions of exposure. His results agree with those of Kerner, *i.e.*, the south slope is warmer than the north, and the east than the west (although the differences are less than in Kerner's results), and the difference is greater the steeper the slope. C. Eser computed the radiation intensities for the different exposures (*ibid.*, VII., 1884, 100-118). Eser found the evaporation from the ground greatest on the south side; then, successively, on the east, west, and north.

In the case of nature, the diurnal range of cloudiness also comes into play in this process. See A. Bühler: "Einfluss der Exposition und des Neigungswinkels auf die Temperatur des Bodens," *Mittheil. schweiz. Centralanstalt f. d. forstliche Versuchswesen*, IV., 1895 [*M.Z.*, XIII., 1896 (23)-(24)].

The differences are slight in forests. Under a dense cover of leaves, the soil temperature is 5°-10°, and on some days as much as 16°, lower than in the open.

fog, and the afternoons are clearer than the mornings. In summer, on the other hand, the forenoons in the mountains are clearer than the afternoons, because ascending currents of air cause cloudiness in the afternoon, and even showers and thunderstorms. The records obtained by means of sunshine recorders have shown that among mountains the noon hour and the early afternoon hours have the most sunshine in winter. The maximum frequency of sunshine is displaced into the forenoon hours as summer comes on, and the sunshine begins to decrease again at 11, or even at 10 A.M., occasionally reaching a slight secondary maximum again late in the afternoon.

The maximum temperature, 20·8°, is reached on September 4 in the Inn valley with an exposure of E. 67° S.; in the Gschnitz valley, with an exposure of E. 76° S., a maximum of 15° is reached on August 25. The lowest temperature, 3·3°, in the Inn valley, was noted on March 5, with an exposure N. 7° E.; and in the Gschnitz valley, – 0·8°, on February 23, with an exposure N. 13° E. The annual ranges are 17·5° and 15·8°.

In the opinion of A. von Kerner, the fact that in the northern Alps the red beech grows to the greatest altitude on the southeastern slopes, while the pine usually grows to the greatest altitude on the southwestern slopes, is explained by these conditions of soil temperature. The beech prefers a warm, dry soil, while the pine prefers a moist soil with a more uniform temperature.

On the mountains of Java, according to Junghuhn, the western slopes below 2400 m., which is the average upper cloud limit, are much damper and cooler than the eastern slopes, because the sun cannot shine upon the former owing to the dense mass of clouds which develops about noon every day. This diurnal march of cloudiness is both extremely regular, and also very clearly defined, within the tropics, at least during a portion of the year. Hence the phenomenon above mentioned, to which Junghuhn has called attention, is probably not limited only to the mountains of Java.

At Innsbruck, the difference between the soil temperatures on the southern and on the northern slopes at a depth of 0·8 m. is over 4° in summer, and must be considerably greater on the surface. For this reason the upper limits of plants, as well as the snow-line, must be considerably higher on the southern than on the northern slopes. The marked difference of exposure and of slope on mountain sides, by affecting the insolation and the air temperature, produce considerable variations of the local climate, even when the places have the same height above sea-level, and are near together. Southern slopes are

much more favourably situated than level ground, for even the low sun of winter may warm them well. Northern slopes, on the other hand, are much less favourably placed than level ground. As a whole, the average value of insolation upon a unit surface of a mountain region is somewhat smaller than that upon a surface of the same area which is level, because in the former case a given pencil of rays is distributed over a larger surface. In this statement no account is taken of the more intense insolation at greater altitudes, which again modifies the result in favour of mountain regions.

CHAPTER XIV.

AIR TEMPERATURE.

Vertical decrease of temperature.—The increase in intensity of insolation with increasing altitude above sea-level, referred to in Chapter XIII., seems contradictory to the vertical decrease in the air temperature. De Saussure, who made so many discoveries in the field of terrestrial physics, was the first to make detailed observations of the rate of decrease of temperature with increasing altitude above sea-level on mountains. In July, 1778, he spent two weeks at an elevation of 3405 m. on the Col du Géant, and while there made an extended series of meteorological and physical observations. Simultaneous observations at Chamonix and Geneva gave the following mean temperatures and rates of vertical decrease of temperature :—

VERTICAL TEMPERATURE GRADIENTS: GENEVA-CHAMONIX-COL DU GÉANT (DE SAUSSURE).

Station.	Altitude (meters).	Mean Temperature.	Decrease of Temperature per 100 meters.	Difference of Altitude for a change of Temperature of 1°
Col du Géant,	3405	2°·5		
			0°·66	150 m.
Chamonix,	1080	17·9		
			0·54	180 m.
Geneva,	400	21·6		

The average vertical decrease of temperature was 0·63° for every 100 meters. It appears further from these observations, and it has since been found to be a general rule, that the decrease of temperature during a gradual ascent up a valley, or on a plateau (*e.g.*, Geneva-Chamonix), is slower than that on a mountain (*e.g.*, Chamonix-Col du Géant).

The table on page 245 gives some examples of the average rate of vertical decrease of temperature in different mountain regions. Except in the first case, and in the case before the last, in the first section of the table, the mean annual temperatures are taken as the basis of the computation.

Weighting Humboldt's rate one-half, because it is based on few observations, and giving the weight of 1 to the rates in the Himalayas, which are really sub-tropical, we have, as a general average for the tropics, a rate of decrease of temperature of 0·56° per 100 m.

From the second section of the following table the general mean for extra-tropical regions, as far as latitude 60° N., is seen to be 0·57°, which agrees with the general mean for the tropics. If we take into account only the average values and the annual means, and disregard local departures, it may be said that the same vertical decrease of temperature prevails in mountain regions from the equator to about latitude 60° N., and averages 0·57° for every 100 m. Even if local peculiarities were to be taken into account, it would be seen that fluctuations between 0·5° and 0·8° would occur in these figures, but these variations show no relation to latitude. Therefore, in the light of our present knowledge of this subject, a dependence of the rate of vertical decrease of temperature upon latitude must be disclaimed.

Effect of topography and of exposure upon the vertical decrease of temperature.—The effect of different exposures is very considerable. In the northern hemisphere the decrease of temperature is more rapid on the southern than on the northern side of the mountains. Thus, in the Swiss Alps, for example, according to Hirsch, the vertical decrease of temperature on the southern side is 0·69°, while on the northern side it is 0·55°; in the eastern Alps it is 0·51° on the north, and 0·60° on the south; in the Erzgebirge it is 0·55° on the north, and 0·63° on the south.

The difference in temperature between a valley and a mountain top rising above it is greater than that between two neighbouring valleys whose altitudes differ as much as do those of the valley and the mountain top. The summits of isolated mountains have lower mean temperatures the more isolated these mountains are, and the smaller their mass is. The vertical decrease of temperature is least rapid in plateau-like mountain districts, and especially on the gradually rising uplands of moderate height which constitute the main masses of continents. In these cases, the vertical decrease of temperature sometimes wholly disappears up to a height of a few hundred meters, and its rate cannot be accurately determined. The determination of the

TABLES SHOWING RATES OF VERTICAL TEMPERATURE DECREASE IN 100 METERS.

I.—Tropical Mountains.

	°
Andes of Colombia and Mexico (Humboldt),	0·53
Andes of South America between lat. 11° N. and 5° S. (Boussingault),	0·57
Andes of Colombia (Hann),	0·51
Andes of Quito (Hann),	0·54
N.W. Himalayas (Blanford),	0·56
N.W. Himalayas with Tibet (Hill),	0·51
Central Himalayas (Blanford),	0·52
Nilgiri Hills (Hann),[1]	0·62
Ceylon (Hann),[2]	0·65
Java[3] (Batavia—Pangerango),	0·56
Island of Hong Kong,	0·57

[1] Dodabetta, 2643 m., 11·8°; Utakamand, 2283 m., 13·3°; Koteghur and Wellington, 1874 m., 16·4°; Coimbatore, 452 m., 25·2°; sea-level (Schlagintweit), 28·0°. These give the equation, $t_h = 27{\cdot}9 - 0{\cdot}62\ h$.

[2] Newara Eliya, 1875 m., 15·1°; Kandi, Badulla, 590 m., 23·5°; Colombo, Batticaloa, Galle, Ratnapoora, Hambantota, 16 m., 27·2° give the equation $t_h = 27{\cdot}3 - 0{\cdot}65\ h$.

[3] Simultaneous hourly observations in May, at Batavia and on the summit of Pangerango.

II.—Extra-tropical Mountains.

	°
Siebengebirge[1] (Bischof),	0·56
Erzgebirge[1] (Reich),	0·52
Erzgebirge (Hann),	0·59
Harz (Hann),	0·58
Alps (Hann, Hirsch, Weilenmann),	0·58
Pyrenees; Pic du Midi (Hann),	0·55
Central and Southern Italy (Lugli),	0·58
Monte Cavo, Rome,	0·55
Etna, Catania,	0·58
Siebenbürgen[2] (Reissenberger),	0·48
Blue Mts., New South Wales[3] (Hann),	0·51
Caucasus and Armenia (Wild),	0·45
Mt. Washington, N.H. (Hann),	0·55
Pike's Peak, Colorado[4] (Hann),	0·63
Colfax, Summit;[4] California (Hann),	0·75
Ben Nevis,	0·67
Near Christiania, lat. 60° N. (Mohn),	0·55

[1] Based on soil temperatures.

[2] Hermannstadt, compared with two stations in the Transylvanian Alps. The latter stations are farther south, and their angular elevation is very slight. Furthermore, Hermannstadt is locally very cold in winter.

[3] Mt. Victoria (1064 m.), compared with Windsor; Bodalla, Albury, Cooma, Kiandra, 1414 m.

[4] Pike's Peak, 4308 m., −7·1°, Colorado Springs, 1830 m., 8·5°, Denver, 1606 m., 9·8°, give $t_h = 19{\cdot}9° - 0{\cdot}63°\ h$. As the plateau is abnormally warm, the summit station gives a very rapid decrease of temperature. Similarly, the valleys of California are abnormally warm.

rate is very difficult even in the case of mountains on a plateau. On the Rauhe Alp, the author found a vertical decrease of temperature of 0·44° in 100 m.; for Wurtemberg as a whole, Schoder obtained a rate of 0·50°; for the Deccan, Schlagintweit found a rate of 0·43°. Further, the stations on the northern and southern sides of the Caucasus, as compared with some stations in the passes among these mountains, give the very slow rate of decrease of temperature of 0·45° in 100 m. In the case of the general continental elevations, the vertical decrease of temperature is probably nearer 0·4° than 0·5° in 100 m.

If the mean temperatures of summit stations at different altitudes are compared with one another, or with the mean temperatures of stations which are freely exposed on steep slopes, the results obtained agree very well with one another; but this rate of vertical decrease of temperature is much more rapid than that noted in the preceding paragraph, and agrees very closely with that in the free air. Thus the summit of the Schafberg (1780 m.), compared with the summit of the Sonnblick (3100 m.), gives a rate of 0·61°; the latter station and Kolm Saigurn give a rate of 0·65°; Ben Nevis and Fort William give 0·67°; Pike's Peak and Colorado Springs give 0·64°. The actual decrease of temperature may therefore probably be taken as 0·65° per 100 m.

Seasonal variations in the rate of vertical decrease of temperature in extra-tropical latitudes.—The cause of the difference in the rate of decrease of temperature between valleys and mountain tops, as compared with that between mountain slopes and mountain tops, is the local cooling of the valleys in winter, especially in climates where the ground is covered with snow in the colder months. Thus the rate of vertical decrease of temperature becomes very slow in winter, and there results a marked annual periodicity in this rate. This annual period in central Europe is seen in the following figures:—

SEASONAL VERTICAL DECREASE OF TEMPERATURE IN CENTRAL EUROPE (PER 100 METERS).

	N. Lat.	Winter.	Spring.	Summer.	Autumn.	Year.
	°	°	°	°	°	°
Harz, - - - - -	52·0	0·43	0·67	0·69	0·51	0·58
Erzgebirge, - - -	50·5	0·43	0·67	0·68	0·58	0·59
Switzerland, - - -	47·0	0·45	0·67	0·73	0·52	0·58
Eastern Alps— (North Side), - -	47·4	0·34	0·60	0·62	0·57	0·51
Eastern Alps— (South Side), - -	46·2	0·50	0·66	0·67	0·57	0·60
Carinthia, - - -	46·7	0·26	0·57	0·58	0·42	0·46

The vertical decrease of temperature in winter is much slower than in summer. It is slowest in Carinthia, where the pressures and winds are the most continental in character. In winter one must ascend, on the average, 220 m. in order to have the temperature fall 1°; in spring, 150 m.; in summer, 140 m.; in autumn, 190 m.; and in the mean for the year, 170 m.

In mountainous regions the rate of vertical temperature decrease is least rapid in December, when the nights are longest and when there is the most favourable opportunity for nocturnal radiation. The most rapid rate occurs in spring and early summer, especially in May and June. Woeikof has shown in detail that this variation is connected with the retreat of the snow cover to greater altitudes in the spring. When the valleys and the highlands are free from snow up to a certain altitude, there is a sudden change in the rate of vertical decrease of temperature as the temporary snow-line is reached; for at and above that point, the radiant energy received from the sun is expended in melting the snow, and the air temperature can therefore rise but little; while farther below the snow-line both the earth's surface and the air can be rapidly warmed. In April, for example, the snow-line at Graubünden is at an altitude of 1000 m., while Chur, at 600 m., is already free from snow. The differences in temperature between Chur and Churwalden (1210 m.) are 4·2° in March, when there is snow at both stations; 4·6° in April, when there is snow at the upper station; and 4·3° in May, when there is no snow at either station. In June the snow-line has retreated to nearly 2000 m.; Sils (1810 m.) is free from snow, but the Julier Pass (2240 m.) is still snow-covered, although the snow is gone in July. The differences in temperature are as follows: May, 2·5°, with snow at both places; June, 3·7°, with snow at the upper station only; and July, 2·6°, when both places are free from snow. The month when the vertical decrease is most rapid therefore depends upon the altitude of the upper station. The greater the altitude of that station above sea-level, the further towards the summer is the time of the maximum rate of decrease displaced.

There are, furthermore, certain general reasons why the most rapid vertical decrease of temperature should occur in spring and early in summer on the earth's surface, as well as in the free air. The results of the meteorological observations made during recent balloon ascents show that the free air at altitudes of 2000-3000 m. is cooler in spring than later on in midsummer. The difference of temperature between the summit of the Schafberg (1780 m.) and the Sonnblick (3100 m.) is greatest in May, namely, 8·9°, which gives a rate of 0·67° in 100 m., although

there is snow on the ground at both stations; and, in this case, there is no sudden change in the rate at the time when the summit of the Schafberg becomes free from snow. The annual variation in the rate of vertical decrease of temperature in the eastern Alps has been somewhat carefully determined by the author, who has found that the slowest rate comes on December 28, when it is 0·33° per 100 m.; and that the most rapid rate comes on May 14, when it is 0·66°. The stations in the passes of the Caucasus, at about 2000 m. above sea-level, show a decrease of temperature of 0·30°, in winter, and 0·57°, in April, in every 100 m. Mt. Washington, in New Hampshire, U.S.A., in latitude 44·3° N., at an altitude of 1914 m. gives 0·43°, in winter, and 0·62°, in summer. Observations made at Christiania[1] gave, for the five winter months, a decrease in temperature of only 0·18°; while the rate for the five summer months was 0·88°. In general, the amplitude in the annual march of the rate of vertical decrease of temperature seems to increase with latitude—a fact which is readily explained.

The temperatures of mountain summit stations when compared with one another, or with the temperatures of stations on mountain sides, where no stagnant, cold air locally lowers the temperature in winter, show a slight annual variation in the rate of vertical decrease of temperature. They also show remarkably accordant values for this rate. For example, the rate between Ben Nevis (latitude, 56·8° N.; altitude 1343 m.) and Fort William (9 m.), is 0·59° per 100 m. in January; 0·79 in May; and 0·67° for the year. The rate between the Sonnblick (latitude, 47·1° N.; altitude, 3106 m.), and the Kolm Saigurn (1600 m.), and the Schmittenhöhe (1950 m.) is 0·53° per 100 m. in January; 0·75° in June; and 0·65 for the year. Between Pike's Peak (latitude, 38·6° N.; altitude, 4308 m.) and Colorado Springs (1840 m.) the rate is 0·52° in January; 0·74° in May-June; and 0·64° for the year. The seasonal means of these values, which are not controlled by local influences, are as follows: winter, 0·58°; spring, 0·71°; summer, 0·70°; autumn, 0·61°.

Seasonal variations in the rate of vertical decrease of temperature in the tropics are slight, and are chiefly dependent upon the changes from rainy to dry seasons, and *vice versa*. The vertical temperature gradient is, in general, somewhat more rapid during the rainy season, and this is probably related to the increase of rainfall and of cloudiness with increase of altitude. The differences between the rainy and the dry sides of mountains are much greater than the seasonal variations,

[1] H. Mohn: "Temperatur in und um Christiania und Wärmeabnahme mit der Höhe daselbst," *Z.f.M.*, IX., 1874, 97-106.

the vertical temperature gradient being much more rapid on the dry side than on the rainy. An example from India illustrates this point.

VERTICAL DECREASE OF TEMPERATURE ON RAINY AND ON DRY SIDES OF MOUNTAINS IN INDIA.

	Ceylon.	Nilgiri Hills.
Rainy (Windward) Side, - -	0·55°	0·56°
Dry (Leeward) Side, - -	0·80	0·71

On the lee side, it is the contrast in temperature between the dry, sunny lowlands and the cloudy and rainy mountain summits which causes the rapid decrease of temperature.

A case which is the reverse of the preceding one is noted by Hill, in the northwestern Himalayas, where the decrease of temperature is 0·76° per 100 m. in January, while it is only 0·41° in July. This anomaly is explained by the fact that the stations at the southern base of the northwestern Himalayas, which are very rainy in July, are compared with the stations in southern Tibet, which are dry, and are therefore relatively very warm in summer. In winter, on the contrary, the lowlands are relatively too warm. To a less marked degree, similar conditions are also met with in extra-tropical mountains. A still greater difference of temperature between mountain summits and the valleys lying to leeward, which occurs from time to time during foehn winds, will later be discussed in detail.

The dependence of the vertical temperature gradient upon the state of the sky in the case of extra-tropical mountains has been studied in detail by Süring. The gradient is less rapid in clear weather, and more rapid in cloudy weather, especially in winter. The mean annual rate is 0·32° per 100 m. in clear weather; and 0·64° in cloudy weather. In clear weather there is always a tendency toward an increase of temperature upward in the morning. This is the so-called *inversion of temperature*, and may reach 500 m. in summer, and much higher in winter.[1]

High-level isotherms. Isothermal surface of 0°.—If we imagine planes drawn through all places having the same temperature in a mountainous region, we shall have a series of isothermal surfaces. If the points of intersection of these surfaces with the mountain sides are

[1] R. J. Süring: *Die vertikale Temperaturabnahme in Gebirgsgegenden in ihrer Abhängigkeit von der Bewölkung*, 8vo, Leipzig, 1890, pp. 34. [*M.Z.*, VII., 1890, (65)-(66).]

connected by lines, these lines are high-level isotherms, such as those which Schlagintweit has drawn for the Alps and for the Himalayas. These isothermal surfaces as a rule have their greatest elevation above sea-level at the equator, and descend towards the poles. Furthermore, as the annual range of temperature is almost zero at the equator, but increases towards the pole, the poleward slope of these isothermal surfaces must be much steeper in winter than in summer. The annual rise of the isothermal surfaces from winter to summer increases with increasing latitude, and is also greater in a continental climate than in a marine climate at the same latitude. In order to gain a clear conception of these relations, it will be best to examine somewhat closely the annual change in altitude of a given isothermal surface, *e.g.*, that of 0°. For the purpose of this consideration we shall confine ourselves to the cases of mountains for which temperature observations are available up to considerable altitudes, so that the altitude at which the temperature of freezing is met with can be stated with some accuracy.

ALTITUDE ABOVE SEA-LEVEL OF THE ISOTHERM OF 0°
(IN METERS).

Mountains.	Latitude.	January.	July.	Year.
	°			
Andes at Quito, - - - -	Equator.	5100	5100	5100
N.W. Himalayas, - - - -	32·0	2800	5700	4700
Mt. Etna, - - - - - -	37·7	1900	3980	2950
Pike's Peak, - - - - - -	38·6	1150	4970	3200
Pic du Midi (Pyrenees), - - -	42·9	1350	3940	2480
Tauern (Eastern Alps), - - -	47·0	0	3200	2050
Ben Nevis (Scotland), - - -	56·8	640	2000	1250

The altitude above sea-level of the isothermal surface of 0° decreases 5100 m. between the equator and latitude 47° N. in winter. In summer it decreases 3000 m. between the equator and latitude 57° N.; and, at that time of year, it nowhere reaches sea-level in the northern hemisphere. It is probably between 300 and 400 m. above sea-level in the neighbourhood of the pole. In the southern hemisphere, however, the isothermal surface of 0° reaches sea-level as far from the pole as latitude 65° S., even in summer. The annual change in altitude of the isotherms is very great in the northwestern Himalayas, being almost 3000 m. A comparison of Mt. Etna with Pike's Peak, in almost the same latitude, is interesting as showing the slight variations in altitude with the

change of season in a marine climate, and the very great variations in a continental climate.

In the eastern Alps, a district for which the author has made careful determinations of the altitude of this isotherm for every month, it is found that the isotherm is nearest sea-level on January 7, at 280 m.; and farthest above sea-level on August 5, at 3550 m. If we take the cases of the northern and the southern sides separately, the altitude is found to be 80 m. on the northern side in January; and 550 m. on the southern side. In summer there is scarcely any difference in the altitudes on the two sides. The altitude of the isothermal surface of 0° increases most rapidly at the beginning of May, when the rise is at the rate of 22 m. a day; it decreases most rapidly at the beginning of November, at the rate of 38 m. a day. The rise in spring is thus seen to be much slower than the fall in autumn.[1]

The observations on the summit of the Sonnblick furnish us with a means of determining, with some accuracy, the temperature on the Glocknergipfel, at an altitude of 3800 m. Its winter temperature is found to be −17°; its summer temperature, −5°; its mean annual, −11°. It is interesting to note that the probable summer temperature of the Glocknergipfel is the same as the observed mean winter temperature of Kolm Saigurn, at 1600 m. The isotherm of about −5° therefore rises from 1600 m. in winter to 3800 m. in summer. The July temperature at the top of Mont Blanc (4810 m.) is shown by observation to be −8°; the mean annual temperature is probably −14·5°. This is the temperature of Arctic North America at latitude 70° N.; but a midsummer temperature of −8° would probably not be found anywhere at sea-level, even at the South Pole. The temperature on the summit of Chimborazo (6300 m.) may be estimated at −6·5°,[2] which is the same as the mean annual temperature of the Sonnblick (3100 m.), in latitude 47° N. The isothermal surface of −6° (annual mean) therefore decreases in altitude 3200 m. between the equator and latitude 47° N., and, in the longitudes of central Europe, reaches sea-level at about latitude 76° N. In north-eastern Russia, sea-level is reached as far south as latitude 68° N.

On the summit of Chimborazo, Whymper observed a temperature of −6·1° in January, and −8·1° in July. On Cotopaxi (5960 m.), Whymper made a reading of −8·4°; Wolf, however, found but −2·0°. On Kilimanjaro (4000-5000 m.), Meyer found temperatures between −11° and −14° at night, with a strong north-west wind, while the temperatures by day were above freezing.

The minimum temperatures observed on mountain tops are as follows:—Pike's Peak (4308 m.), −39·4°; Mt. Washington (1914 m), −45·6°; Sonnblick (3100 m.), during eight winters, −34·6°; Pic du Midi (2860 m.), during fifteen winters, −34·8°; Mont Blanc, during the winter of 1894-95, −43·0°; Mt. Ararat (5100 m.), during two winters, −50° and −40°. In the last three cases the

[1] J. Hann: "Die Temperaturverhältnisse der oesterreichischen Alpenländer," *S.W.A.*, XCII., 1885, 78-82. The height of the isotherm of 0° in the free air over Berlin has been discussed by Berson in *Wissenschaftliche Luftfahrten*, III., 1890, 101.

[2] J. Hann: "Klima von Quito," *Zeitschr. f. Erdkunde* (Berlin), 1895, 122.

readings were taken from minimum thermometers which had been left on the mountain summits during the winter.

Inversions of temperature.—During clear nights, especially during the winter in middle and higher latitudes, it is observed that, in calm weather, the valleys are colder than the slopes and the summits of the enclosing mountains, up to a certain height. Observations made in different places have shown that even in the free air, as over a plain, there is a vertical increase of temperature during clear calm nights throughout the year, but especially in winter, when the ground is covered with snow. This increase of temperature upward reaches altitudes of at least 300 m., and is rapid in the lower strata, but slower farther up. At Montpellier, Martins found that this increase of temperature upward, on clear nights, was at the rate of about 1° in every 10 m. In the lower strata the rate was about 0·7° in 2 m., and even more than that in individual cases. Thus, in the case of trees of 6 m. and more in height, the difference in temperature between the crown and the surface of the ground may easily exceed 2°. This explains the fact that on frosty nights the tree tops may remain unharmed, while the lower branches, as well as the shrubs, are frostbitten. If the sky is cloudy, and there is considerable wind, this phenomenon either does not occur at all, or else it is but very faintly developed.

The hourly observations made on the Eiffel Tower (300 m.), in Paris, have shown that there is an increase of temperature upward throughout the year between midnight and 4 A.M.; this increase being greatest in autumn and least in spring. In the afternoon, however, between 12 M. and 4 P.M., there is a very rapid decrease of temperature upward. An earlier series of observations, made at a height of 39 m. above the ground on the Great Pagoda, in Kew Gardens, gave a similar result. For the year, the mean temperature minima on the Pagoda were 0·3° higher, and the mean temperature maxima were 0·6° lower. In summer, the mean maxima aloft were 1° lower. On perfectly clear days, the differences of the higher station, as compared with the lower, were as follows :—

DIFFERENCE OF TEMPERATURE BETWEEN TOP AND BASE OF PAGODA IN KEW GARDENS.

	9 A.M. °	3 P.M. °	9 P.M. °	Mean 9 A.M. and 9 P.M. °
Winter, - -	0·3	−0·2	0·5	0·4
Summer, - -	−0·1	−0·7	0·3	0·1

It appeared that, in clear weather in winter, it was 0·4° warmer at 39 m. than on the ground. In foggy weather, both in winter and summer, it was more than 1° warmer at the greater elevation.[1]

Inversions of temperature explained.—The cause of this anomalous vertical distribution of temperature is clearly to be found in the nocturnal radiation from the surface of the earth. The resulting cooling of the ground likewise affects the atmosphere which lies on, or near the surface, and as cold air is heavier than warm, it follows that when the air is calm the coldest layers lie nearest the ground. The upper strata cool but little, because the radiation from the air itself is much less than that from the earth's surface, and from the vegetation which may cover that surface.[2]

These conditions, however, certainly give no direct explanation of the increase of temperature at night, from a valley to the enclosing hill sides and hill tops; because the increase of temperature with altitude, just described as due to the cooling of the air in contact with the earth's surface when the latter has been cooled by radiation, does not extend to so great a height as from a valley to a mountain side. Furthermore, the mountain slopes and summits likewise themselves cool through losing heat by radiation, and we should expect that this latter cooling would be even greater than that of the valley bottom; because, as has already been noted, radiation increases with increasing altitude. The cause is, nevertheless, the same in both cases. Inversions of temperature result from nocturnal radiation, and from the arrangement of masses of air of different temperatures in strata, according to

[1] A. Angot: "Sur la Décroissance de la Température dans l'Air avec la Hauteur," *Comptes Rendus*, CXV., 1892, 1270-1273. R. H. Scott: "Results of Observations made at the Pagoda, Royal Gardens, Kew, and elsewhere, to determine the Influence of Height on Thermometric Readings, on Vapour-Tension and on Humidity,' *Quart. Weather Rept.*, N.S., 1881, Part I., Appendix III. See also G. J. Symons: "Note on the Establishment and First Results of Simultaneous Thermometric and Hygrometric Observations at Heights of 4 and 170 Feet, and of Siemens' Electrical Thermometer at 260 Feet above the Ground," *Proc. Roy. Soc.*, XXXV., 1883, 310-319; M. H. Carlier: "Observations méteorologiques faites à Saint Martin de Hinx, Bayonne, 1875" (*Z.f.M.*, XI., 1876, 123-125). A. Sieberg: "Temperaturumkehrungen mit der Höhe zwischen Aachen und dem Aussichtsthurm im Aachener Walde," *M.Z.*, XVIII., 1901, 33-34; and J. Hann: "Einige Ergebnisse der Temperaturbeobachtungen auf dem Strassburger Münsterthurm," *ibid.*, 211-216.

[2] Cf. also the interesting results obtained on two balloon voyages, made from Munich on summer nights. S. Finsterwalder and L. Sohncke: "Einige Ergebnisse wissenschaftlicher Fahrten des Münchener Vereines für Luftschiffahrt," *M.Z.*, XI., 1894, 361-376, and especially p. 374.

their specific gravity, the coldest layer remaining at the lowest level as long as the atmosphere is calm enough to maintain this arrangement of successive strata.

Low temperature of valleys in winter.—Let us now examine the nocturnal increase of temperature up to a height of a few hundred meters. This phenomenon occurs in hilly or mountainous districts on clear, calm nights, throughout the year. Wells made the statement that there is more radiation in valleys than on open plains, and there has been much discussion concerning the accuracy of this view. Koosen therefore undertook a special series of observations with reference to this point, at his residence in Weesenstein (near Dresden, Germany), which is surrounded by hills from 30 to 60 m. high; and his results confirmed the accuracy of Wells's statement.[1]

The radiation on calm, clear nights was considerably greater on the level valley floor than on the plateau lying above it. The temperature of objects close to the ground, and also of the air itself up to a given height, was often 4° – 5° below that of the plateau temperature. On the morning of September 23, 1862, the minimum thermometer, at a height of 6 m. above the ground, showed a temperature of –2·5° and hydrangeas, fuchsias, canna, caladium, and araucaria excelsa had their blossoms and leaves frozen. The leaves of grape-vines and the young shoots of acacias and roses were very much damaged; while on the plateau, 30 to 60 m. above the valley, not a single plant—not even the dahlia—was injured by frost.

In explaining this phenomenon, which is noted in spring, and still more strikingly in autumn, it must be remembered that the loss of heat by radiation begins in the valleys from one to two hours earlier in the afternoon than on the uplands, and lasts as much longer in the morning. In going from a valley on to an upland shortly after sunset, in summer, it may be noticed that all the plants in the valley are covered with dew, while the plants on the upland are still quite dry. The relative intensity of radiation is, however, probably about the same in the valley as on the upland, because radiation is most active toward the zenith, and decreases rapidly toward the horizon. There is, moreover, the further consideration that calms prevail in valleys much more frequently than on plains, or exposed uplands, and wind to a great extent equalises the differences of temperature which result from nocturnal radiation, and thus keeps the vegetation from being injured. It must also be remembered that the layers of air which have been cooled by radiation from the ground flow down the slopes into the valley, and collect there, while the air on the mountain slopes is continually renewed by a supply of warmer air.

[1] Koosen: *Pogg. Ann.*, CXVII., 1862, 611.

The water vapour of the air is frequently condensed into thin sheets of fog in the cold, stagnant air of valley bottoms. In going down into a valley which is well covered with vegetation on a warm, calm summer evening, the sensation of increasing dampness and coolness bears witness to the distribution of temperature above noted, even without the aid of any instrumental observation.

Relation of inversions of temperature to vegetation.—The considerations of the preceding paragraphs make it clear that mountain slopes and hill tops have the climatic advantage of less cold at night than is found in the neighbouring valleys. This favourable characteristic is strikingly illustrated in the cultivation of the more delicate plants of economic importance. Toward the polar limits of the zone occupied by each particular species of plant, there is found to be greater immunity from damage by frost on hill sides than in valley bottoms. In the province of San Paulo, in Brazil (latitude, 20-25° S., altitude, 500-800 m.), the coffee plantations are laid out on the hills only, and never in the valley bottoms. The reason for this is that frost occurs in the depressions between the hills, but very rarely on the hills themselves. Houses built on the sides or tops of hills have the advantage of less humidity, as well as of higher temperatures at night.

Pammer, the meteorological observer at Pisino, in Istria, says that this station lies in a hollow, in which the mountain torrent, Foiba, disappears in an underground cavern. This valley is almost always filled with fog on calm nights, and to this fog is due the high relative humidity of Pisino. The temperatures of the city itself are such that the olive and the fig do not thrive there, while these trees are found all over the hills surrounding the city, at a height of 100 to 200 m. above it. The temperature on the hills must, therefore, be considerably higher than in the city itself. A. von Kerner also remarks that the vegetation at the bottom of the basins in the limestone district of the Karst leads to the conclusion that the temperature is lower there than in the surrounding country.

Some instructive examples of the effects of inversions of temperature have recently been given by Stow.[1] At the end of May, 1894, there was frost in the interior of England. After the frost, the ash trees in the valley bottoms were completely frost-bitten; those which grew somewhat higher on the slopes were frost-bitten below their crowns, but the crowns were green; those which grew on still higher ground had their lower halves frozen, but their upper halves were green; while the ashes which were growing on the hills surrounding the valley were not injured in the slightest. During the severe winter of 1879-80,

[1] F. W. Stow: "A Natural Thermometer," *Quart. Journ. Roy. Met. Soc.*, XXI., 1895, 214-215.

Colonel Ward observed a minimum temperature of $-37°$ in a valley bottom at Rossinière, in southwestern Switzerland; while the minimum at his house, only 30 m. higher, was but $-12°$.

A typical description of the frost zones which depend upon the local topography in the Alleghany Mountains has been given by Silas M'Dowell, of Franklin, Macon county, North Carolina.[1] After a spell of pleasant spring weather, in March and April, terminated by a few days' rain, there comes a clearing off with cold northwest winds, "leaving the atmosphere of a pure indigo tint." If the wind subsides at night, the following morning shows a heavy hoar frost; plants and all kinds of fruit trees are killed, "and the landscape, clothed in verdure the day before, now looks dark and dreary." Under these conditions the beautiful phenomenon of the "verdant zone" or "thermal belt" exhibits itself upon the sides of the mountains. "It begins at about 100 m. above the valley, traversing them in a perfectly horizontal line throughout their entire length, like a vast green ribbon upon a black ground. Its breadth is 400 ft. (120 m.) vertical height, and from that under, according to the degree of the angle of the mountain with the plane of the horizon. Vegetation of all kinds within the limits of this zone is untouched by frost; and such is its protective influence that the Isabella, the most tender of all our native grapes, has not failed to produce abundant crops in twenty-six consecutive years; nor has fruit of any kind ever been known within these lines to be frost-killed, though there have been instances where it has been so from a severe freeze. The lines are sometimes so sharply drawn that one-half of a shrub may be frost-killed, while the other half is unaffected."

The height of the frostless belt above the valleys varies in different cases. Observations on this point were made by M'Dowell in Macon county, North Carolina, "which is traversed by the beautiful valley of the Little Tennessee River, lying 2000 feet (600 m.) above tide-water. Here, when the thermometer is down to 26° F. ($-3{\cdot}5°$ C.), the frost reaches about 300 feet (100 m.) vertical height. A small river, having its source in a high plateau 1900 feet (580 m.) above this, runs down into this valley, breaking through three mountain barriers, and consequently making three short valleys, including the plateau, rising one above the other, each of which has its own vernal zone traversing the hillsides that enclose them, the first of which takes a much lower range than that of the lower valley, and each taking a lower as the valleys

[1] J. W. Chickering, Jr.: "Thermal Belts," *Amer. Met. Journ.*, I., 1884-85, 213-218.

mount higher in the atmosphere, and in the highest one the range of the belt is not more than 100 feet (30 m.) above the common level of the plateau, a beautiful level height containing 6000 acres of land, and lying 3900 feet (1200 m.) above tide water."[1]

A similar belt is found along the eastern slope of the Tryon Mountain range, in Polk county, North Carolina. The differences in temperature between the top and the base of these mountains and the thermal zone which lies between, are so great that they are perceived without the aid of a thermometer. On summer nights, these differences amount to 3° – 6°, and on winter nights, to 8° – 11°.

It is often noticed that while the temperature falls very rapidly on the slopes and summits of mountains early in the evening, this condition of things is soon reversed, and the temperature may even rise again near, or after, midnight. The temperatures registered at Kolm Saigurn have often shown this. A similar case is cited by Clayton as having occurred at Blue Hill, Mass., during the clear sky and dry air of a summer anticyclone (August 15-23, 1886). The temperature on the summit rose steadily during the night of August 22-23, and was higher at sunrise than in the evening; while at the base the temperature fell continuously till sunrise, at which time it was 6° lower than at the summit.[2]

The foresters in the Thuringian forest speak of "frost-holes." Similarly, in the Rhine valley, frosts commonly occur locally on lowlands.

The less severe cold of the slopes and the summits of mountains as compared with that of the valley bottoms is especially noticeable during the colder spells of winter. Even the mean winter minima are lower in the valleys than on neighbouring higher land from which the cold air can easily drain away. In the Alps, the coldest valleys are those which are protected on the west against the warm winds that frequently blow with considerable velocity from that direction, and in these valleys the conditions are therefore favourable for collecting masses of air which have been cooled by radiation. In valleys of this kind, the average vertical distribution of temperature shows an increase of temperature with altitude at the beginning, or even in the middle of

[1] Cf. A. Kerner: "Die Entstehung relative hohen Lufttemperaturen in der Mittelhöhe der Thalbecken der Alpen im Spätherbste und Winter," *Z.f.M.*, XI., 1876, 1-13.

[2] H. H. Clayton: "Cause of a Recent Period of Cool Weather in New England," *Science*, VIII., 1886, 233, 281-282. See also, regarding the climatic characteristics of hills, "Observations upon the Climate of Crowborough Hill, Sussex," 1885, and *Symons's Met. Mag.*, XII., 1878, 188-189. This paper gives examples of warm nights on the uplands of Hampshire.

the winter. In connection with this, it must be remembered that this winter cold is not usually imported by cold north or northeast winds, but is locally produced by radiation under the clear sky of the long nights, especially if the ground is covered with snow. The cooling of the air under these conditions normally proceeds from below upward, and the cold strata have at first very little vertical extent. Hence the distribution of low temperatures in winter is dependent upon all the circumstances which affect radiation and the unimpeded accumulation of masses of cold air. Thus Bevers, in the valley of the upper Engadine, has a January temperature of $-10 \cdot 4°$, at an elevation of 1715 m. If the ascent is continued to the Julier Pass, at 2244 m., a January mean of $-8 \cdot 8°$ is there met with, and, on the Rigi Kulm (1784 m.), which has about the same altitude as Bevers, the January mean is only $-5 \cdot 1°$. The respective mean winter minima of the places just named show the same relation; thus, Bevers has $-26 \cdot 9°$; Julier Pass, $-23 \cdot 9°$; Rigi Kulm, $-18 \cdot 9°$; Grächen (1632 m.), in the upper Valais, situated on a mountain slope, has a January mean of $-4 \cdot 4°$; while Davos, in a valley at the same altitude (1650 m.), has $-7 \cdot 3°$. The mean winter minima are $-17 \cdot 3°$ and $-24 \cdot 7°$, respectively.[1]

In the valleys of Carinthia, the vertical increase of temperature in winter is so well known that it has given rise to a proverb which may be freely translated as follows:—

> "If you climb in winter up a stair,
> One less coat you will need to wear."

Places at an intermediate altitude are the warmest; but the mean temperature at 2000 m. above sea-level is higher than that of the valley bottom. The mean annual minimum at Klagenfurt (440 m.) is $-21 \cdot 7°$; that of St. Paul, in the Lavant valley (390 m.), is also $-21 \cdot 7°$; while at Hüttenberg (780 m.), the corresponding reading is $-14 \cdot 8°$; and at Lölling (1100 m.), $-14 \cdot 7°$. Both of these stations are on the western slopes of the Sau Alp. At Tröpolach (590 m.), the mean annual minimum is $-24 \cdot 1°$; and on the Obir (2040 m.), only $-21 \cdot 0°$.

Air currents during winter anticyclones in the Alps.—During the clear, calm weather that accompanies winter anticyclones, systems of slowly descending, cold air currents are developed over the whole of the Alpine district. The direction and the velocity of these currents are controlled by the inequalities of the surface. Like streams of water, these currents of air tend to unite in the ravines and side valleys, and to converge toward the main valleys, in which they continue their

[1] All these temperatures relate to the same period, 1864-1871, but are not to be considered normal means.

descent after the manner of rivers in their channels. The air which thus flows down from the slopes must be replaced by a new supply. Hence air descends from aloft upon the mountain tops and slopes, and this air is not only initially warmer than that which was cooled by contact with the cold ground, but is also warmed by compression during its descent. It is a general principle, to the discussion of which we shall return later, that air which descends from a lower pressure aloft to a higher pressure nearer sea-level, without being cooled by any external cause, is itself warmed. This may be illustrated by the artificial compression of air in an ordinary air pump. Thus it happens that extraordinarily high temperatures occasionally prevail on mountain tops for several days at a time in midwinter, while the temperatures in the valleys are considerably below freezing. Under these conditions, a layer of fog several hundred meters in thickness, which results from the marked cooling of the lower strata of the atmosphere, usually covers the valleys and the lowlands; while the air on the slopes and the summits of the mountains is very dry. The low temperatures are not limited to any fixed altitude, but all the valleys are cold, and all the slopes and the summits are warm.[1]

The months of November, December, and the first half of January, are the most favourable for the development of this phenomenon, because the nights are then the longest.

Inversions of temperature in mountain regions in winter.—The eastern Alps furnish numerous examples of an increase of temperature with increase of altitude during the winter, especially when the upper stations are on mountain slopes. As there is a decrease of temperature with increasing altitude in summer, it follows that, with increasing altitude, the climate becomes more equable. The valley of the Ahrn, which unites with the valley of the Puster at Bruneck, furnishes some excellent examples of this, as is seen in the following table:—

INVERSIONS OF TEMPERATURE IN THE AHRN VALLEY.

Station.	Altitude (Meters).	January.	July.	Difference.
		°	°	°
Bruneck, - -	825	−6·8	18·1	24·9
Steinhaus, - -	1050	−5·8	15·9	21·7
St. Peter, - -	1360	−4·6	14·7	19·3
Prettau, - -	1440	−5·6	13·1	18·7

[1] Note the interesting and important observations made by A. von Kerner, at Innsbruck and in its vicinity, during an inversion of temperature of this kind (*Z.f.M.*, XI., 1876, 1-13).

Toblach (1252 m.), which is built on a very flat alluvial fan, from which there is a good opportunity for the cold air to drain away, has a temperature of – 7·3° in January and – 6·1° as a winter mean; while Gratsch (1175 m.), lying but 80 m. below, in a hollow, has – 9·4° in January and – 7·8° as a mean for the winter. The latter station is therefore about 2° colder. The increase of temperature with increase of altitude is most strikingly seen in Carinthia. Starting, for example, from Klagenfurt, and ascending the western slopes of the great Sau Alp along the Görschitz valley, the winter temperatures are found to be distributed as follows :—

INVERSIONS OF TEMPERATURE IN CARINTHIA.

Station.	Altitude (Meters).	January.	Winter.
		°	°
Klagenfurt, - - -	440	– 6·2	– 4·6
Eberstein, - - -	570	– 4·2	– 3·3
Hüttenberg, - - -	780	– 3·1	– 2·3
Lölling, valley, - -	840	– 2·5	– 1·6
Lölling, mountain house,	1100	– 1·9	– 1·3
Stelzing, - - -	1410	– 3·7	– 3·2

The Lölling mountain house on the southern slope of the Erzberg, at Hüttenberg, has an extraordinarily mild winter temperature, which is even milder than that of Graz, 750 m. lower. January, at the former station, is 1·5° warmer than at Graz, and 4·3° warmer than at Klagenfurt. The camp at an altitude of 1180 m., on the western slope of the Kor Alp, has almost as high a winter temperature, namely, – 1·9°, the January mean being – 2·2°; while St. Andrä, which lies below in the Lavant valley, has – 4·6° and – 6·1°, respectively. If the January temperatures at an altitude of about 1000 m., on the western slopes of the Kor Alp and the Sau Alp, are reduced to sea-level, the result is 2·5°; while the valley bottoms, at 400 m., give – 4·4°. The slopes are thus seen to be 7° warmer.

Low temperatures of valleys the result of nocturnal radiation.— The abnormal cold of the valley bottoms is the result of nocturnal radiation, and the mild temperatures of the mountain slopes at intermediate altitudes are the result of the descent and the mechanical warming of the air from aloft. That these contrasted conditions of temperature are not, as is often supposed, due to the more favourable

situation of the higher land with regard to sunshine, is shown by the fact that the differences in temperature between the mountain slopes and the valley bottoms are greatest at night, when the sun is below the horizon. The mean difference in temperature between Lölling mountain house and Klagenfurt, 660 m. lower, as determined by observations made during six winters, was 3·9° at 6 A.M. before sunrise, and only 2·3° at 2 P.M. The vertical increase of temperature per 100 m. was therefore 0·59° at the end of the night, and only 0·35° under the influence of the afternoon sun. Furthermore, the air at Lölling is dry in the morning and the evening, and moist at noon, which is further evidence of a warming of the air that descends along the mountain slopes at the time when the cold is greatest in the valley. Even in the southern Tyrol, in the valley of the Etsch, between Bozen and Ala, an increase of temperature upward is noted in cold winters, there being an accumulation of cold air over the lower ground, for which the valley of the Etsch does not supply a sufficient means of drainage.[1]

Inversions of temperature during winter anticyclones in Europe.— Within recent years the occurrence of long spells of anticyclonic conditions over central Europe, notably in December, 1879, January, 1880, and January and February, 1881, resulted in some very remarkable inversions of temperature over the whole district of the Alps, and even in the central mountains of Germany. A few examples will serve as illustrations.

MEAN TEMPERATURES OF DECEMBER, 1879, IN CARINTHIA.

Altitude (meters), - - -	450	580	830	1200	2040
Mean Temperature, - - -	° −13·3	° −10·4	° −8·8	° −6·9	° −9·4
Minimum Temperature, - -	−25·6	−21·1	−19·3	−18·5	−24·4

[1] The vertical distribution of temperature during a spell of high pressure and clear weather in Pinzgau in January, 1867, has been fully discussed by the author. The mean temperature at 850 m. was −8·3°. From that altitude—the level of the valley bottoms—the temperature increased at first at the rate of 0·9° per 100 m. as far as 1700 m. above sea-level. The initial temperature of −8·3° was not reached again until an altitude of 2520 m. was attained; and thence upward the temperature decreased. A stratum 1670 m. in thickness was therefore warmer than the valley bottom. (J. Hann: "Resultate des ersten Jahrganges der meteorologischen Beobachtungen auf dem Sonnblick," *S. W. A.*, XCVII., 2a, 1888, 5-38.) See also Billwiller: "Vertikale Temperaturvertheilung innerhalb barometrischer Maximalgebiete in verschiedenen Jahreszeiten," *Z.f.M.*, XVI., 1881, 89-94; also G. Greim: "Temperaturumkehrungen im vorderen Odenwalde," *M.Z.*, IX., 1892, 417-421, and Eliot on inversions in India, *Journ. Asiat. Soc. Bengal*, LIX., 1890, Part II., No. 1, 1-50.

The vertical increase of temperature was most rapid during the time that the centre of the anticyclone lay over the Alps themselves, *i.e.*, during the 13 days from December 16 to 28. The mean temperatures during this period were as follows :—

MEAN TEMPERATURES AND MEAN CLOUDINESS AT FOUR STATIONS IN CENTRAL EUROPE FOR DECEMBER 16-28, 1879.

Station.	Altitude (meters).	7 A.M.	2 P.M.	9 P.M.	Mean.	Mean Cloudiness (0 – 10).
Klagenfurt, - - -	440	° – 19·1	° – 13·0	° – 16·4	° – 16·2	3·2
Summit of Obir, - -	2040	– 5·9	– 1·2	– 5·5	– 4·5	1·7
Ischl, - - - -	467	– 13·7	– 7·3	– 13·0	– 11·8	1·6
Schafberg, - - -	1776	– 0·1	0·6	– 1·3	– 0·5	0·7

Thus, while the temperature was truly Siberian in its intensity at Klagenfurt, the temperature on the Obir was mild, and on the summit of the Schafberg it was not far from the freezing-point, even at night.[1] It appears, further, as has been noted above, that the difference of temperature was greatest at 7 A.M., *i.e.*, that the greater altitudes are relatively the warmest at that time. This is direct evidence that the whole phenomenon was an effect of radiation and of the slow descent of the cold air into the valleys. During this same period, inversions of temperature were also strikingly shown in the valley of the Po, according to Cantoni. The mean temperatures and the absolute minima for a period of 60 days, beginning with the second decade

[1] The warmth of the higher strata is strikingly shown in the following figures, which give the mean temperatures at 7 A.M., *i.e.*, before sunrise, for the period December 20-24, 1879.

Station.	Altitude (meters).	Temperature.
Altstätten, - -	480	° – 12·7
Trogen, - - -	890	– 3·9
Gäbris, - - -	1250	3·4
Rigi, - - - -	1790	1·2

Altstätten is on the flood-plain of the Rhine.

of December, 1879, and ending with the first decade of February, 1880, were as follows:—

MEAN TEMPERATURES IN NORTHERN ITALY, DECEMBER, 1879—FEBRUARY, 1880.

Station, - -	Alessandria.	Padua.	Milan.	Varese.
Altitude (Meters), -	98	98	147	862
Mean Temperature,	° − 8·5	° − 7·6	° − 5·7	° −1·0
Absolute Minimum,	−17·0	−14·0	−10·5	−9·4

It is thus seen that the stations which had the highest temperatures were those at the greatest elevations and nearest the Alps, while the lower stations were the coldest. Alessandria is nearer than Padua to the axis of the Po valley, along which the cold air slowly descended, and therefore the former station was the colder. The low winter temperatures of northern Italy are explained in the same way as are those of the Carinthian valley. Both of these districts act as catch-basins for the air which has been cooled by conduction and radiation to the cold ground, and which flows down into them from the surrounding higher land. In the case of the plain of northern Italy, the mountain ranges on the north, west, and south form a barrier against the prevailing winds, and hence make it possible for the cold air to stagnate over the lowlands to leeward.

Relative humidity during inversions of temperature.—The following observations made during the same cold spell may serve to illustrate the dryness of the warm air aloft.

TEMPERATURE, RELATIVE HUMIDITY AND CLOUDINESS ON THE PUY DE DOME AND AT CLERMONT FOR DECEMBER 20-28, 1879 (6 A.M.).

Station.	Altitude (meters).	Temperature.	Humidity (per cent.).	Cloudiness (per cent.).
Puy de Dôme, -	1470	° 3·8	38	13
Clermont,[1] - -	390	−13·2	91	7

During the anticyclone of October 8-9, 1900, over the northern Alps, the relative humidity on the Sonnblick (3106 m.) fell to 17 per cent., and on the Zugspitze (2964 m.) to 9 per cent. The observer on the

[1] At the base of the Puy de Dôme.

Zugspitze reported an exceptional clearness of the higher strata of the atmosphere during this period, while heavy fogs hung over the lowlands.[1]

In the region of the great permanent anticyclone and of the winter "cold pole" of eastern Siberia, inversions of temperature are probably normal phenomena, and the severe winter cold of the valleys is replaced by higher temperatures at greater altitudes.[2]

Inversions of temperature and human habitations.—The inversions of temperature above described, which result in giving the summits and slopes of mountains higher temperatures than prevail on the lowlands, have, in many cases, been the determining factor in the location of human settlements. Thus in the Alps, many farms and villages are built on the mountain sides, often at considerable distances from the fields and pastures which belong to them, rather than on the level valley floors which, in many respects, are much more convenient. Kerner has given a picturesque description of the conditions which make it easy for the traveller of the present day to see why the peasants who built these dwellings selected the higher elevations as building sites. The Swiss have learned by experience that the mountain sides have far more favourable temperature conditions in late autumn and in winter than the lowlands. During one of the calm, clear spells of late autumn the traveller who spends a few days at one of these farm-houses on the steep mountain side may there breathe air which has the mildness of summer; he may see the green fields still decked with autumn flowers, and may watch the sheep grazing in the fields, while down below, in the valley, the ground is already frozen hard by the frost, the trees are leafless, and all the activities of plant life have long ceased.

Löwl calls attention to the fact that settlements in the Alps are not, as a rule, found on valley floors and ancient lake beaches, but rather on alluvial fans, terraces, and similar topographic forms. The motive in this selection of building sites is evidently to avoid the cold, which is charac-

[1] J. Hann: "Das Barometer-Maximum vom 8 und 9 Oktober, 1900, und die Witterung auf den Hochgipfeln der Nordalpen," *M.Z.*, XVII., 1900, 565-567.

[2] Further particulars in this connection may be found in the author's discussion: "Die Temperaturverhältnisse der oesterreichischen Alpenländer," III., *S.W.A.*, XCII., 2, 1885, 89-105, and *Zeitschr. des Deutsch. und Oesterreich. Alpenvereins*, 1886, 51.

In regard to the temperatures and the humidities aloft in anticyclones, see J. Hann: "Ueber das Barometermaximum vom November, 1889, in Mittel-Europa nebst Bemerkungen über die Barometermaxima im Allgemeinen," *D.W.A.*, LVII., 1890, 401-424.

teristic of the valley bottoms during the winter and at night, as well as to escape the fog and dampness which are common in those localities.[1]

Some observations, made at Neukirchen, in Pinzgau, have shown how considerable these differences in temperature really are. The temperature at Neukirchen itself in January, 1888, was −5·8°; twenty paces lower it was −6·5°, and on the floor of the valley, only 40 or 50 m. lower, it was −9·3°. Separating the clear and the cloudy days, the temperatures were as follows, the mean cloudiness being given in brackets.

TEMPERATURES AT NEUKIRCHEN IN JANUARY, 1888.

	Clear Days (1·5).			Cloudy Days (9·2).		
Jan., 1888,	9 A.M.	2 P.M.	8·30 P.M.	9 A.M.	2 P.M.	8.30 P.M.
	°	°	°	°	°	°
Station, -	−15·0	− 7·4	−11·8	−2·5	0·7	−1·6
Valley floor, (45 m. lower)	−21·1	−10·7	−17·6	−3·2	0·2	−2·6

The minimum temperature at the station was −21·2°, and on the valley floor, −29·5° (9 A.M.). These observations seem to offer a sufficient explanation for the avoidance of the valley bottoms as building sites.[2]

Inversions of temperature in New England have been considered by Davis.[3] When the records of adjacent hill and valley stations during winter anticyclones are compared, it is found that the lowest nocturnal minima always occur in the valleys, while the hills have less extreme cold. On a higher summit, like Mt. Washington (1914 m.), the temperatures are distinctly moderate. Early on the morning of Jan. 19, 1887, after a clear quiet night, the observer at Waterbury, Conn., found a gradual rise of temperature from −31·7°, at the lowest point in the valley, to −17·8° at an elevation of 60 m. on the enclosing hills. At Lexington, Mass., one thermometer 2·4 m. above the ground read −26°, while another on the same building, but 9 m. above the ground, read −21·7°. On the morning of Dec. 27, 1884, when the pressure was above the mean and the wind was light, with a clear sky, the temperature on Mt. Washington was −8·9°, while that on the neighbouring lowlands ranged from −23° to −31°.

Causes of the vertical decrease of temperature.—The facts of the vertical decrease of temperature have now been considered. It remains

[1] F. Löwl: "Siedlungsarten in den Hochalpen," *Forschungen zur Deutschen Landes- und Volks-kunde*, II., No. 6, Stuttgart, 1888.

[2] See also *M.Z.*, IV., 1887, 184-185; V., 1888, 148-149; VI., 1889, 146; X., 1893, 190-192.

[3] W. M. Davis: "Types of New England Weather," *Ann. Astron. Obs. Harv. Coll.*, XXI., Pt. II., 1890, 119.

to give an explanation of them. It was noted at the close of the chapter on Solar Climate (Chapter VI.), that the storage of heat at the bottom of the atmosphere results from the peculiar behaviour of this atmosphere toward solar radiation. This process has been called *selective absorption*. The radiations of shorter wave-length, including the luminous rays, are less absorbed, but more scattered, while the radiations of greater wave-length—the invisible infra-red rays—suffer a greater selective absorption, and are to some extent altogether prevented from reaching the earth's surface. As solar radiation is very rich in rays of such wave-length as are readily transmitted by the atmosphere, a large proportion of this radiation is available for warming the earth's surface. On the other hand, the "heat rays" which are emitted from the earth's surface are to a very considerable extent absorbed by the atmosphere, because this is non-luminous radiation, of long wave-length, in the extreme infra-red portion of the spectrum. Thus it is seen that the radiation from the sun passes to the earth's surface through the atmosphere more freely than the non-luminous radiation from the earth passes out again through the atmosphere. In this way, the atmosphere helps to store up heat at the earth's surface, and this process of storage is naturally most effective in the lower strata, which are the densest and contain the most impurities, and is least effective in the rare, dry, and clean air of greater altitudes.[1]

[1] Trabert explains this process as follows. The temperature remains constant if the gain of heat (Q), equals the loss by radiation, which, following Stefan, may be set down as $\mathrm{a}T^4$, T being the so-called *absolute temperature*, whose zero is at $-273°$ C. We therefore have $Q=\mathrm{a}T^4$, and the resulting constant temperature is

$$T=(Q\div\mathrm{a})^{\frac{1}{4}} \quad \text{.......(1)}$$

But the earth's atmosphere must also be considered in this problem; and this atmosphere transmits one portion of Q better than another portion. If α represents the coefficient of transmission of the "luminous rays," Ql; and β represents the coefficient of transmission of the "non-luminous rays," Qd, the amount originally transmitted through the atmosphere from the sun is $\alpha Ql+\beta Qd$. But, on the way out through the atmosphere from the earth's surface, the expression for the amount now transmitted is assumed to be $\beta\mathrm{a}T^4$, because the radiation from the earth is non-luminous. These two values must again be equal when the temperature becomes constant. By simultaneously adding βQl to, and subtracting it from, the first expression—a process which does not alter the value of the latter—we obtain

$$\alpha Ql+\beta Qd+\beta Ql-\beta Ql=\beta\mathrm{a}T^4$$

or

$$(\alpha-\beta)Ql+\beta(Qd+Ql)=(\alpha-\beta)Ql+\beta Q=\beta\mathrm{a}T^4$$

whence

$$T=\left\{(Q\div\mathrm{a})+\frac{\alpha-\beta}{\beta}\,\frac{Ql}{\mathrm{a}}\right\}^{\frac{1}{4}} \quad \text{.......(2)}$$

The temperature, T, is therefore now higher than it was previously in (1), in consequence of the intervention of the atmosphere, for the sum of the right side is

Therefore the thinner the atmospheric envelope, the less the effect of the atmosphere, and the lower the temperature of bodies within it, which are then exposed to a freer receipt and loss of radiant energy. The mean temperature of the air must be distinguished from that of the surface which the atmosphere protects. The air temperature decreases with increase of altitude, in spite of the increase in the intensity of solar radiation with the corresponding decrease in the vertical thickness of the absorbing envelope. Exception must be made, however, in the case of an elevated zone of incipient absorption, for which the vertical temperature gradient is nearly zero.[1]

Effect of conduction and radiation to and from the earth's surface.— The foregoing facts would not, however, explain the higher temperatures of plateaus as compared with mountain summits of equal height. To make the explanation complete, a further circumstance still remains to be taken into account, viz., the direct effect of the earth's surface, which has been warmed by insolation, upon the atmosphere which rests upon that surface. The ground is warmer than the air during the day, because the land, being a dark, opaque body, is a much better absorber of radiant energy than the atmosphere itself. Furthermore, the warming of the ground is concentrated near the surface, because the land does not transmit the radiant energy to any great depth. The ground communicates its heat to the lowest layers of the air which lie upon it, in part directly by conduction, and in part by radiation outward from the ground, the coarse waves from the ground being absorbed by the air. Thus the ground becomes a very effective source of heat for the air which is near it, and it is clear that the temperature of the air over a plateau, or over an extended mountain region, must be higher than that of the air around an isolated mountain peak, or than that of the free air at the same altitude.

The fact that the atmosphere is almost always in motion, makes it impossible for well-warmed masses of air to accumulate around isolated

increased. The temperature is higher, the greater the coefficient of transmission, α, for the luminous rays, as compared with the coefficient of transmission, β, for the non-luminous rays. Thus the temperature, T, at the earth's surface increases if α increases with respect to β.

As a matter of fact, there is very little difference between α and β for the two regions of the solar spectrum indicated, but a very great difference if α signifies the transmission of solar rays and β that of telluric radiation, for α is then much larger than β. Equation (2) should therefore be written

$$T=\sqrt[4]{\frac{Q}{\mathrm{a}}\times\frac{\alpha}{\beta}}.$$

[1] F. W. Very: "Atmospheric Radiation," 123.

mountains. The air around such mountains is quickly carried away again, and moreover the mountain mass itself, which warms the air by conduction and radiation, being very limited in area, is not a very effective source of heat. During the winter, and at night, on the other hand, the ground cools by radiation more than the air. Hence the temperature of the air over a plateau at such times is lower than that of the air on a mountain top, or at the same altitude in the free air.

Changes of temperature in ascending and descending air currents.—The rate of decrease of temperature with increase of altitude depends upon the interaction of several causes. In winter, and especially at night, the earth's surface, so far from being a source of heat for the lower air, is a source of cold. Hence, as has already been seen, inversions of temperature occur in calm weather. If calm clear weather were to prevail throughout the winter, inversions of temperature would be the normal condition.

If the air is in motion, there is abundant opportunity for the lower and the upper air to become mixed, and in mountainous regions the air is carried upward along the slopes. Under these conditions, there is a rapid decrease of temperature upward, because rising air cools at the rate of 1° for every 100 m. of ascent. Let us imagine a mass of dry air enclosed in a balloon made of some non-conducting material, and so constructed that the balloon will expand to any desired extent. Under these conditions the enclosed air will always assume a volume corresponding to the changes in pressure outside the balloon. The pressure within and without will, therefore, always be the same. Now, if the balloon be allowed to rise, in consequence of a higher temperature of the air within than that of the air outside, the ascending air will continually expand as a result of the continued decrease of the pressure with increasing altitude. At the same time, the temperature of the air inside the balloon will fall at the rate of 1° for every 100 m. of ascent, as we learn from the mechanical theory of heat. This loss of heat is the thermal equivalent of the work done by the air in increasing its volume against the pressure of the surrounding air outside the balloon. This loss of heat can be made good at any time; because, if the balloon were to be drawn down to the earth's surface again by a cord, the enclosed air would warm at precisely the same rate as that at which it cooled during the ascent. Thus the air inside the balloon would reach the ground with the same excess of temperature as that with which it began its ascent. In this process of regaining the initial temperature, however, the air in the balloon would also be reduced to its original volume again, because of the increasing pressure without. The general

law may therefore be stated as follows: Ascending masses of air cool at the rate of 1° for every 100 m. of ascent, and descending masses of air warm at the same rate. This rate applies only in the case of air which is not cloudy, *i.e.*, whose temperature is above the dew-point.

Stable, unstable and indifferent equilibrium.—If we imagine the upper and lower strata of the atmosphere so thoroughly mixed that every particle of air has been several times to the upper limit of the atmosphere and back again, the vertical decrease of temperature would be at the rate of 1° in every 100 m. Under these conditions of vertical temperature gradient, both ascending and descending masses of air would find their own temperatures at all altitudes. In other words, both ascending and descending air would everywhere be in equilibrium, and would nowhere have any tendency either to rise or to fall. This is the condition of *indifferent equilibrium.*

When the vertical decrease of temperature is more rapid than at the rate of 1° in 100 m., a mass of air induced to rise would, at every altitude, find a temperature lower than its own at that altitude, and would therefore tend to continue its ascent to the top of the atmosphere. On the other hand, cold masses of air could also come down from aloft to the earth's surface under these conditions. Theoretically, air does not ascend voluntarily, of its own initiative, until the rate of vertical decrease of temperature amounts to 3·4° in 100 m.[1] It is not till these conditions of vertical temperature gradient are reached that the specific gravity of the lower layers of air is the same as that of the higher. This is the condition of *unstable equilibrium.*

On the other hand, with a vertical decrease of temperature of less than 1° in 100 m., which is actually the usual condition, ascending movements of the air must soon cease, and descending masses of cold air are so rapidly warmed that they cannot reach the earth's surface as a cold blast. This is the condition of *stable equilibrium* in the atmosphere.

In this connection the low temperatures which have thus far been observed at great heights in the atmosphere may be noted. Barral and Bixio found a temperature of – 39·7° at an altitude of 7000 m. above Paris during their balloon ascent of July 27, 1850, immediately after a thunderstorm. Yet even this abnormally cold mass of air would have been warmed 70° if it had been brought directly down to the earth's surface, and therefore would have had a temperature of 30° at sea-level. On May 11, 1894, Berson, in the balloon *Phönix,* observed a temperature of – 36·5° at an altitude of 7930 m.; and on December 4,

[1] J. Hann: *Lehrbuch der Meteorologie,* 1901, 752-754.

the same undaunted aeronaut, with the same balloon, reached a height of 9150 m., where he found a temperature of $-47 \cdot 9°$. A temperature of $-67°$ was recorded on one of the voyages of the unmanned balloon *Cirrus* at an altitude of 18,000 m., and on the voyage of the balloon *l'Aerophile* on October 20, 1895, at 15,500 m., $-70°$ was recorded. It is the rapid warming of descending air which protects the earth's surface against invasions of cold air from aloft.

The condition of *indifferent equilibrium* would occur in the atmosphere if the only source of heat were at the earth's surface, because then the temperature which each particle of air would have at every altitude would be due solely to the heat of the earth's surface. It may be assumed that this condition of so-called *convectional temperature equilibrium* exists in the atmosphere of the sun. In the case of the earth's atmosphere, however, the rate of vertical decrease of temperature is much slower than this condition would involve. It is, in fact, on the average only one-half as rapid. We must therefore conclude from this fact, if from no other, that the warmed surface of the earth cannot be the only source of the heat of the atmosphere. It has already been explained that the atmosphere itself absorbs some of the radiant energy received from the sun. The upper strata also receive a considerable increase of heat directly from the sun.

Retarded rate of cooling in cloudy ascending currents.—The condensation of water vapour in ascending currents of air is a second important source of heat for the higher strata of the atmosphere. Moist air contains a much greater quantity of heat than dry air at the same temperature. If, for example, a kilogram of dry air, at a temperature of 25°, is cooled to 0°, it must suffer a loss of a quantity of heat $0 \cdot 238°$[1] $\times 25 = 5 \cdot 95$ calories. This loss of temperature may be brought about by an ascent of about 2500 m. If, on the other hand, we take a kilogram of air at the same temperature saturated with water vapour, the cooling to 0° results in the condensation of about 16 grams of water vapour, whose latent heat of condensation is equivalent to $16 \times 0 \cdot 6 = 9 \cdot 6$ units of heat. The kilogram of saturated air must therefore lose $9 \cdot 6 + 5 \cdot 9 = 15 \cdot 5$ kilogram-calories. The heat lost in the case of the moist air whose temperature is reduced to freezing is therefore nearly three times as great as in the case of the dry air. Hence, it may be seen that, as soon as condensation begins, ascending moist air cools much more slowly than ascending dry air.

Rates of cooling with ascent in moist air of different temperatures.—The following table shows the rates of decrease of temperature in

[1] Specific heat of air.

each 100 m. of ascent, in the case of saturated air at different temperatures. The rate of decrease of temperature is naturally slower the higher the temperature of the saturated air, *i.e.*, the more water vapour is contained in each kilogram of moist air. The pressure is included in this table for the reason that the quantity of vapour in a kilogram of saturated air increases with increase of altitude, *i.e.*, with decreasing pressure.

RATES OF DECREASE OF TEMPERATURE PER 100 METERS IN SATURATED AIR AT DIFFERENT TEMPERATURES.

Initial Pressure.	Initial Temperatures.					Altitude (Meters).
	-10·0°	0·0°	10·0°	20·0°	30·0°	
760 mm.	0·76	0·63	0·54	0·45	0·38	0
500 mm.	0·68	0·55	0·46	0·38	—	3360

The rate of vertical decrease of temperature must be much slower in a moist atmosphere, in which vertical air movements are developed as the result of the greater warming of the lower strata of the air, than in a dry atmosphere. Since the air is not, at all times and in all places, saturated with water vapour, the decrease of temperature during the daytime is more rapid in the lower strata, but is slower above the altitude at which condensation most frequently takes place. In so far as the vertical decrease of temperature depends upon the condensation of the water vapour, the rate should be slower in warmer and damper climates, at least above a certain altitude, than in colder and drier climates; and, other things being equal, the rate should be more rapid in dry climates than in moist climates. The rate should therefore be slower over oceans and along sea coasts than over continents. The results of observation accord, in general, with this hypothesis. The mean annual vertical temperature gradient in the tropics averages about the same as that in higher latitudes. This results from the fact that the more rapid decrease of temperature in the summer of the higher latitudes, which is in accordance with the demands of theory, is compensated by the slower decrease of temperature in winter, when, because of the increased radiation, the surface of the cold ground cools the lower strata of the atmosphere.

Vertical temperature gradients on mountains.—The vertical temperature gradient in mountainous districts is most rapid when the weather

is windy or stormy, because it then most closely resembles the temperature gradient of ascending air currents. It is slowest, on the other hand, during calm weather, because there is then no vertical movement of the whole mass of the atmosphere, and the process of heating is a general and uniform one. It is true that, during calm clear summer weather, high valleys and mountain slopes and summits do have a somewhat higher temperature than is found in the free air at a corresponding altitude, but this results from the fact that the warm ground heats the air by conduction and radiation. Furthermore, our thermometers indicate a somewhat higher temperature than the air really has. In general, however, the greater cold which is felt on mountains during high winds is not to be explained on physiological grounds alone, but is actually demonstrable by the thermometer. It results from currents of air ascending the mountain sides. The mountain tops may then be colder than the free air at the same altitude.

CHAPTER XV.

ANNUAL AND DIURNAL MARCH OF TEMPERATURE IN MOUNTAIN CLIMATES

Decrease of annual range of temperature with increase of altitude. —The amount of the annual range of temperature generally decreases with increasing elevation above sea-level, the decrease being slight in the tropics, but of considerable importance in middle and in higher latitudes. The difference in temperature between the warmest and the coldest months at certain stations in Ceylon and India is as follows [1] :—

ANNUAL RANGES OF TEMPERATURE AT DIFFERENT ALTITUDES IN INDIA AND CEYLON.

At Sea-level.		In the Interior.	
Colombo (rainy west coast), -	2·0°	Kandy (522 m.), - - -	2·7°
Batticaloa (dry east coast). -	3·8	Newara Eliya (1875 m.), -	2·1
Southern India, Nilgiri Hills.			
Coimbatore (452 m.), - -	5·1	Utakamand (2283 m.), - -	4·9
Kotagiri, Wellington (1874 m.) -	5·1	Dodabetta (2643 m.), - -	3·8

In these cases the annual range of temperature is very small even at sea-level, and shows scarcely any decrease with altitude. When the lower station has a continental climate, the decrease with altitude becomes more marked because the amount of warming is less on

[1] See also A. Woeikof: "Temperatur und Hydrometeore auf dem Agustia Peak in Südindien und am Fusse des Berges," *M.Z.*, XIII., 1896, 405-416.

mountains, while the cloudiness and the rainfall are greater than at lower levels.

The following examples are from Northern India :—Roorkee, 270 m., 18·8°; Chakrata, 2150 m., 15·5°; Simla, 2119 m., 15·8°; Rawalpindi, 503 m., 22·9°; Murree, 2276 m., 18·6°. Calcutta and Goalpara, 118 m., on the other hand, have a range of temperature of 10·3°; Darjeeling, 2107 m., has 12·6°. Further examples are :—St. Helena, Jamestown, 12 m., 5·2°; Longwood, 538 m., 5·1°; Hongkong, Victoria, 13·3°; Victoria Peak, 532 m., 12·2°; Aden, 60 m., 7·2°; Gondar, 2270 m., 5·8°.

It is thus seen that in mountain districts within the tropics the annual range of temperature decreases either very slightly, or not at all, with increase of altitude.

In middle and higher latitudes, where the decrease of temperature with altitude has so marked an annual period, and where this decrease is much slower in winter than in summer, mountain stations at considerable heights have a smaller annual range of temperature variations than do stations on the lowlands. Mountain climates therefore resemble the climate of seacoasts in this respect. The following data illustrate this particular :—

ANNUAL RANGES OF TEMPERATURE AT DIFFERENT ALTITUDES.

United States.

Burlington and Portland,	70 meters.	27·9°
Mt. Washington,	1914 ,,	22 9
Denver,	1610 ,,	24·7
Pike's Peak,	4308 ,,	20·9

Caucasus and Plateau of Armenia.

Vladikavkas-Tiflis,	570 meters.	24·3°
Gudaur,	2160 ,,	22·4
Erivan, Aralykh,	870 ,,	35·1
Alexandropol,	1470 ,,	29·7

Italy and Sicily.

Rome,	50 meters.	18·4°
Monte Cavo,	960 ,,	17·0
Catania,	30 ,,	16·2
Etna,	2947 ,,	10·8

FRANCE.

Clermont, - - - -	390 meters.	18·0°
Puy de Dôme, - - - -	1467 ,,	13·3
Toulouse, - - - -	190 ,,	16·4
Pic du Midi, - - - -	2860 ,,	14·1

ALPS.

Station.	Altitude (meters).	January.	July.	Difference.
Altstätten, - -	460	–1·2°	18·2°	19·4°
Trogen, - - -	880	–1·2	15·9	17·1
Gäbris, - - -	1250	–2·0	13·5	15·5
Rigi, - - -	1790	–4·8	9·7	14·5
Säntis, - - -	2470	–8·1	5·7	13·8
Zell a S., - - -	750	–5·9	16·1	22·0
Bad Gastein, - .	1020	–4·0	14·8	18·8
Kolm Saigurn, - .	1600	–5·4	12·5	17·9
Schafberg, - -	1780	–5·1	9·8	14·9
Schmittenhöhe, -	1950	–7·1	9·0	16·1
Sonnblick, - -	3100	–13·0	1·5	14·5

If we could assume that the annual range of temperature continued to decrease up to the greatest altitudes at the same rate as appears in the foregoing tables, the difference in temperature between the seasons would disappear at about 9500 m. above sea-level. The recent international balloon ascensions in Europe, and the observations carried out by de Bort by means of *ballons-sondes*, have, however, shown that even up to 10,000 m. there may be quite a marked tendency to an annual variation of temperature. Thus the annual march of temperature at great altitudes in the free air, as determined by de Bort on the basis of numerous observations obtained by means of balloons, is as follows :—

ANNUAL RANGE OF TEMPERATURE IN THE FREE AIR (DE BORT).

Altitude.	Minimum.	Maximum.	Range.	Mean.
3 km.	– 11·2° Feb.	2·2° Aug.	13·4°	– 4·5°
5 km.	– 20·8 March.	– 7·6 ,,	13·2	– 14·2
10 km.	– 52·9 ,,	– 43·9 ,,	9·0	– 48·1

A decrease in the annual range of temperature with altitude is distinctly seen in the foregoing data, the range at Paris being 16°. The amount of the range at 10 km. is, however, surprising.[1]

Dependence of annual range of temperature on topography.—We should, nevertheless, be greatly mistaken if we assumed that the decrease in the annual range of temperature with increasing altitude were as regular as the foregoing data might lead us to believe. This regularity is found only when stations in valleys are compared with those on mountain slopes, or when stations of the latter class are compared with one another, for topography is an important factor, as may be seen from the following cases:—

ANNUAL RANGES OF TEMPERATURE AT DIFFERENT ALTITUDES, ILLUSTRATING THE INFLUENCE OF TOPOGRAPHY.

	°		°
Rigi Kulm, 1784 m., - -	14·5	Denver, 1610 m., - -	24·7
Sils and Bevers, 1762 m., -	20·4	St. Louis, 150 m., - -	26·0
Chur, 603 m., - - -	18·6		

Sils and Bevers have a range of temperature which is 6° greater than that at the Rigi Kulm, whose altitude above sea-level is the same, and is also greater than that at Chur, which is 1160 m. below them. Similarly, we find annual ranges of temperature on the plateaus west of the Mississippi Valley which are almost the same as the ranges in the Mississippi Valley itself. Leh, at an altitude of 3517 m. in Tibet, has an annual range of temperature of 25·3°; Peshawer, at an altitude of 390 m., in the same latitude, has a range of only 22°. Plateaus and elevated valleys as a whole show hardly any decrease in the annual range of temperature with increasing altitude. In individual cases they even show an increasing range aloft. This is a natural consequence of the fact that the valleys become abnormally cold during the winter, partly as a result of radiation and partly because of the accumulation of cold air in them; while, on the other hand, they are very well warmed in summer.[2]

[1] See L. T. de Bort: "Sur les Ascensions dans l'Atmosphère d'Enrégistreurs météorologiques portés par des Cerfs-volants," *Comptes Rendus*, CXXIX., 1899, 131-132; "Sur la Température et ses Variations dans l'Atmosphère libre, d'après les Observations de 90 Ballons sondes," *ibid.*, 417-420 (translated by C. Abbe in *Mo. Weather. Rev.*, XXVII., 1899, 411-413); *Wissenschaftliche Luftfahrten*, Berlin, 1900, Vol. III.

[2] Compare the stations in the Ahrn valley, on page 259.

The relatively high temperature of the valley bottoms in summer is primarily due to their favourable situation with reference to insolation, whereby, assuming the sun's altitude to be the same, the eastern slopes of the enclosing mountains are better warmed than the level ground in the morning, and the western slopes in the afternoon. The heat reflected and radiated from the sides of the valleys, as well as the protection afforded against cooling winds, also play a part. The climate of valleys, especially at considerable altitudes above sea-level, is extreme in consequence of increased radiation during the winter, and stronger insolation during the summer in the rarer and drier atmosphere aloft. In a mountainous country the ranges of temperature, therefore, vary greatly with the location of the station in a valley, on a slope, or on a mountain top; and also with its situation on a northern or a southern slope, or in a valley opening toward the east, or the west.

Time of occurrence of maximum and minimum temperatures on mountains.—The annual march of temperature at considerable altitudes on the slopes or the summits of mountains is similar to that in coast climates, both in the decreased difference between the extremes of temperature, and also in the retardation of the time of occurrence of these extremes as compared with stations in the same latitude on lowlands. The time of occurrence of the minimum temperature on mountains, in particular, is delayed until February, and even until March. The differences of temperature between the mountains and the lowlands at their base are greatest in the early spring months. When the winter snow covering has already disappeared from the lowlands, and the ground there is being well warmed by the sun which is high in the heavens, all the insolation at greater altitudes must still be devoted to the task of melting the snow.

The following table illustrates the annual march of temperature in the lower Alpine valleys,[1] in the higher Alpine valleys, and on the mountain summits. The annual march of temperature in the Dalmatian Islands shows how much the conditions there resemble those on elevated mountain summits. So close is this resemblance, indeed, that it may truly be said that the march of temperature on mountain tops approaches that in a purely marine climate.

At the greater altitudes, and in the marine climate as well, April is still cold, and even the temperature of May does not rise far above the annual mean. On the other hand, October is a warm month, and November is much milder than March.

[1] Based on the means for the northern and southern sides of the Alps.

ANNUAL MARCH OF TEMPERATURE IN THE ALPS AND IN THE MARINE CLIMATE OF THE ADRIATIC SEA.

Departure of the Monthly Means from the Annual Means.

	Valleys. 400 meters.	High Valleys. 1900 meters.	Summits. 2400 meters.	Dalmatian Islands. 0 meters.
	°	°	°	°
December, -	−10·4	−8·3	−6·5	−6·6
January, - -	−11·4	−9·0	−6·9	−7·6
February, - -	−8·9	−7·5	−6·7	−7·4
March, - -	−4·9	−5·1	−5·5	−5·7
April, - -	0·7	−0·6	−1·5	−2·0
May, - -	4·7	3·1	2·2	1·8
June, - -	8·7	7·0	5·7	6·2
July, - -	10·5	9·2	8·1	8·5
August, - -	9·8	8·8	8·0	8·4
September, -	6·1	5·9	5·4	5·6
October, - -	0·9	1·4	1·5	1·7
November, -	−5·9	−4·8	−3·9	−3·1
Range, - -	21·9	18·2	15·0	16·1

The annual minimum occurs in the Alpine valleys on January 8, but it does not occur until the middle of January on the summits of the Alps, and still later, on January 22, in the Dalmatian Islands. The annual maximum occurs in the Alpine valleys on July 23; on the summits, on August 2; and in the Dalmatian Islands, on July 30. In the spring, the temperature rises above the annual mean in the lower Alpine valleys on April 15; in the higher valleys, on April 24; on the summits, on May 1; and in the Dalmatian Islands, on May 5. The temperature falls below the annual mean again at these stations on October 19, 23, 24 and 29, respectively. In the Alpine valleys, the temperature remains above the mean during 185 days; but on the summits of the Alps, and in the Dalmatian Islands, it is above the mean for 177 days only, or 8 days less. On the Alpine summits, the temperature rises for 200 days, and falls for 165 days, the fall being, therefore, more rapid than the rise; while at the lower stations, the rise lasts 195 days, and the fall 170 days.[1]

[1] J. Hann: "Temperaturverhältnisse der oesterreichischen Alpenländer," III., *S. W. A.*, XCII., 2, 1885, 38-61.

The diurnal march of temperature on mountains shows both local and general modifications. Local modifications are brought about chiefly by the exposure of a station, which determines whether the hours of sunshine shall be shortened early in the morning or late in the afternoon. Further, local winds, especially the cold winds which blow down many valleys at night, markedly affect the rapid fall of temperature after sunset, and retard the time of occurrence of the minimum temperature. The average diurnal march of temperature at any station in a mountainous region must therefore be discussed only in the light of an accurate knowledge of the local conditions, and not with reference to the general rule which governs the march of temperature at other stations in the same latitude.

Valleys usually have a greater diurnal range of temperature than stations over neighbouring lowlands. The temperature of the air rises rapidly during the day because the valley floor and the surrounding mountain sides are well warmed, and the latter also help to warm the air within the valley. Even the air of the valley wind, which blows up the slopes during the day, has already been somewhat warmed. After sunset, however, the temperature falls rapidly, and cool mountain winds blow down from the higher slopes and out of the cold shady ravines. The air which has been cooled by nocturnal radiation on the mountain slopes spreads out over the valley bottoms, where it stagnates, its temperature being still further reduced as the result of radiation from the valley floor. Valley bottoms, therefore, have a somewhat greater diurnal and annual range of temperature than neighbouring stations on lowlands.

Decrease of diurnal range of temperature with increasing altitude. —The case is different on the sides of mountains, as well as in valleys with a steep descent, in which there is a good opportunity for air drainage. On mountain sides, the nights are much milder, and also less damp, than in the valleys. The days are also cooler, unless it happens that the exposure is such as to increase insolation very considerably. With increasing altitude on mountain summits, the decrease in the diurnal range of temperature is more rapid the more isolated the mountain, and the smaller the mountain mass; in other words, the decrease in the diurnal range of temperature is more rapid the less the air temperature can be affected by the warming and the cooling of the ground. An example of this is found in the following data:—

DIURNAL RANGE OF TEMPERATURE AT DIFFERENT ALTITUDES IN THE ALPS.

JULY-AUGUST.

Station.	Altitude (meters).	Range.
		°
Geneva, - - -	407	10·6
Chamonix, - -	1035	14·2
St. Bernard, - -	2470	4·4
Grands Mulets, -	3010	4·9
Mont Blanc, - -	4810	3·5

The large diurnal range in the enclosed valley of Chamonix, as compared with the more open location of Geneva, is shown in a very striking manner, as is also the increase in the range with increasing altitude. It should be said, however, that the diurnal range of temperature on the summit of Mont Blanc, as observed, is undoubtedly too large, the heat reflected from the surface of the snow probably having considerable effect. Three stations in Japan, all of which are near together, and for which we have simultaneous observations taken every two hours, show the same phenomenon in August, 1891. Nagoya, situated on a plain at an altitude of 15 m. above sea-level, had a diurnal range of 6·8°; Kurosawa, in an enclosed valley at 830 m. above sea-level, had a range of 10·1°; Ontake, a mountain top, 3055 m., had 5·7°. Synchronous observations, made by Kämtz on the Faulhorn (2680 m.), and by Horner at Zürich (480 m.), gave for the month of September a diurnal range of 4·8° at the upper station and 11·7° at the lower. The stations on the plateau and in the valleys of the Rocky Mountains furnish similar examples of the large temperature ranges in valleys.[1] Thus, for example, Sherman, Wyoming, in an open pass, at an altitude of 2530 m., has a mean diurnal range of 9·9°; while Georgetown, Colorado, has 16·4°. The foregoing conditions have been well summarised by Woeikof in the following statement: A *convex* surface (such as a hill or a mountain) *diminishes* the diurnal and the annual range of temperature; while a *concave* surface (such as a valley or other depression) *increases* the diurnal as well as the annual range of temperature.[2]

[1] J. Hann: "Ueber den täglichen Gang des Luftdruckes, der Temperatur, der Feuchtigkeit, Bewölkung und Windstärke auf den Plateaux der Rocky Mountains," *S. W. A.*, LXXXIII., 2, 1881, 484-503; and *Z.f.M.*, XVII., 1882, 31-40.

[2] "*Die Klimate der Erde*, I., Chap. 8.; and "Temperaturänderung mit der Höhe in Bergländern und in der freien Atmosphäre," *M.Z.*, II., 1885, 201-218.

The rapidity of the decrease in the diurnal range of temperature with increasing altitude above the earth's surface has been shown with remarkable distinctness by the hourly observations of air temperature made at different altitudes on the Eiffel Tower in Paris.[1] The open iron-work of which this tower is built affects the air temperature by its own heating, and the consequent radiation of heat, to a much less degree than does even the steepest and most freely exposed mountain summit. Taking the daily march of pressure on mountain summits as a basis, the author has calculated the daily march of temperature in the atmosphere between mountain summits and the adjacent lowlands, and a study of these data, as well as of the observations made on the Eiffel Tower, has led to some interesting results. This study has furnished convincing proof that most mountain summits have shown too large a diurnal range of temperature. The surface of the ground, which is well warmed during the day, raises the observed temperature above that of the free air at the same height; while, on the other hand, nocturnal radiation operates in the reversed direction. This action takes place even when the mountains are wholly covered with snow or with glaciers. The observed temperature on mountain tops can approach the temperature of the free air closely only while the air is in active motion.

[1] The diurnal range of temperature in the lower air depends almost altogether on the diurnal range of temperature of the surface upon which the air rests, *i.e.*, of the ground. The process of warming the air by interchanging ascending and descending air currents is very gradual, and causes a decided retardation in the occurrence of the maximum temperature in the upper air. Thus the maximum of the soil temperature comes at 1 P.M., and that in the upper air between 5 and 6 P.M. The direct warming of the air by insolation is of hardly any importance at all in the case of the lower air as compared with the process just mentioned. Even at Katherinenburg, the temperature of the surface of the ground rises to 27·9° in the 1 P.M. mean for May-August; and this is 8·7° higher than the air temperature. At Nukus, the surface temperature rises at the same time to 52·4°, which is 22·9° higher than the air temperature. The diurnal range is 16·8° at Katherinenburg and 37·4° at Nukus, while the ranges in the air are 8·8° and 14·2°, respectively. This shows very clearly the influence of the daily march of temperature on the surface of the ground upon the diurnal range of temperature in the lower air. As the diurnal period of warming is very short, and is being continually interrupted by the cooling of the surface below the air temperature at night, the diurnal range of temperature cannot reach as far up in the atmosphere as does the annual range. Furthermore, the greatest hourly change in the air temperature during the diurnal period is more than 100 times as great as that during the annual period (Woeikof). The amplitude of the diurnal variations in air temperature must therefore decrease rapidly with increase of altitude, about in a geometrical progression.

Diurnal ranges of temperature at different altitudes.—The two most exposed summit stations in the Alps,[1] and the stations on the Eiffel Tower, show how rapidly the diurnal range of temperature decreases with increasing altitude.

DIURNAL RANGE OF TEMPERATURE AT PARIS, ON THE EIFFEL TOWER, AT KLAGENFURT, ON THE OBIR, AND ON THE SONNBLICK.

	Paris.	Eiffel Tower.			Klagenfurt.	Obir.[3]	Sonnblick.[3]
	18 m.[2]	123 m.[2]	197 m.[2]	302 m.[2]	450 m.	2140 m.	3100 m.
	°	°	°	°	°	°	°
Winter, -	3·6	3·0	2·5	2·0	5·5	0·9	1·1
Spring, - -	6·9	5·8	5·0	4·6	9·1	2·0	2·4
Summer, -	7·7	6·1	5·2	5·2	9·6	3·5	2·2
Autumn, -	6·1	4·7	3 6	2·9	5·7	1·4	1·5
Year, - -	6·1	4·9	4·1	3·7	7·5	1·8	1·7

At a height of 300 m. above Paris, the diurnal range of temperature has decreased nearly one-half. The decrease is very rapid at first, and then slower. Angot[4] concludes that, at an altitude of about 750 m. in winter, and 1150 m. in summer (mean for the year, 900 m.), the diurnal range of temperature in the free air at Paris is reduced to one-tenth of its value at sea-level. The observations on the Obir and on the Sonnblick show that, at 2000-3000 m., the diurnal range of temperature is very small. An estimate of the mean diurnal ranges at different altitudes, based on the foregoing observations, is contained in the following figures :—

MEAN PERIODIC DIURNAL RANGES OF TEMPERATURE AT ALTITUDES OF 0 TO 2600 METERS.

Altitude (meters), - -	0	100	200	300	1700	2600
	°	°	°	°	°	°
Range, - - - - -	7·5	6·4	5·6	5·1	2·0	1·8

[1] The diurnal march of temperature upon the Säntis as determined by a thermometer exposed in a new location, on the summit, has not yet been discussed.

[2] Relative heights of the thermometer above the ground.

[3] See also J. Hann: "Ueber Temperatur des Obir-Gipels (2140 m.), und des Sonnblick-Gipfels (3106 m.)," *S. W.A.*, CVII., 1898; IIa, 537-568 (*M.Z.*, XVI., 1899, 575-576).

[4] A. Angot: "Résumé des Observations météorologiques faites au Bureau Central et à la Tour Eiffel, 1890-1894," *Ann. Bur. Centr. Met. de France*, 1894, I.

Taking the daily march of pressure on the Eiffel Tower and on mountain summits, and using these data to determine the diurnal ranges of temperature in the free air, the resulting values are still smaller for the summer. The author found the following :—

MEAN DIURNAL RANGES OF TEMPERATURE IN SUMMER AT ALTITUDES OF 140 TO 3200 METERS IN THE FREE AIR.

Altitude (meters), - -	140	240	680	840	2000	3200
Range, - - - -	4·3°	3·3°	2·2°	1·7°	1·4°	1·0°

Classifying the observations by clear and by cloudy days, the following ranges are found for the Alps in summer :—

MEAN DIURNAL RANGES OF TEMPERATURE ON CLEAR AND ON CLOUDY DAYS IN THE ALPS IN SUMMER.

Altitude (meters), - -	230	850	1600
Clear, - - - - -	4·0°	2·8°	2·0°
Cloudy, - - - -	2·0	1·0	0·7

Hence it appears that the observed diurnal ranges on the Obir and on the Sonnblick are too large as compared with those in the free air. The temperature observations obtained during kite ascents at Blue Hill Observatory show that the diurnal range of temperature diminishes rapidly with increasing altitude in the free air, and almost disappears on the average at a height of 1000 m. Probably, however, it occasionally extends to altitudes of 2000 m. The average diurnal range (determined by direct observations and computed from the kite meteorograph records at different altitudes, counting from the level of the Valley Station, which is itself 15 m. above sea-level) are as follows :—

DIURNAL RANGES OF TEMPERATURE AT DIFFERENT ALTITUDES, AS DETERMINED BY OBSERVATIONS AT BLUE HILL OBSERVATORY.

	Valley Station.	Base Station.	Summit Station.	Kite.	Kite.
Diurnal Range, -	11·6°	9·9°	9·3°	2·4°	0·17°
Altitude (meters),-	0	49	180	500	1000

The range at the summit of Blue Hill is believed to be somewhat too great because of the effect of the heating and cooling of the surface of the hill upon the temperature of the air.[1]

The time of occurrence of the daily extremes of temperature varies with altitude as well; but in this case local influences play so large a part that this characteristic can hardly be said to vary according to any definite law. The march of temperature on a high mountain top or slope is determined by two factors, viz., the daily march of temperature in the free air, and that resulting from the heating of the surface of the ground at the station itself. In the case of the first factor, the times of the occurrence of the extremes are greatly delayed as compared with those of the earth's surface, the maximum at considerable altitudes in the free air apparently occurring as late as about 5.30 P.M. The march of temperature which is under local control, on the other hand, has its maximum soon after noon. The local temperatures are further dependent upon the marked diurnal variation of cloudiness, especially in summer, the cloudiness increasing rapidly toward noon, and checking the further rise of the surface temperature. This probably explains the discrepancy in the hours at which the daily maximum occurs. Kämtz found the maximum on the Faulhorn at a quarter of an hour after noon; Smits, on Pangerango (2950 m.), found the maximum as early as 11.30 A.M. in May. At Batavia, at the same time, the maximum came at 2 P.M., evidently as the result of the diurnal variation in cloudiness. The observations on the St. Bernard and Theodul Pass show a maximum in summer at 1.15 P.M., and at 1 P.M., respectively.

In Austria-Hungary, on the other hand, the most exposed mountain summit stations have their daily extremes at the following hours:—

HOURS OF MAXIMUM AND MINIMUM CLOUDINESS ON THE OBIR AND ON THE SONNBLICK.

	Winter.	Spring.	Summer.	Autumn.	Year.
Obir, 2140 m.					
Minimum, - - -	6 A.M.	8 A.M.	3.30 A.M.	4 A.M.	4.30 A.M.
Maximum, - - -	2 P.M.	3 P.M.	3 P.M.	2 P.M.	2.30 P.M.
Sonnblick, 3106 m.					
Minimum, - - -	5 A.M.	4.30 A.M.	4.30 A.M.	5 A.M.	4.30 A.M.
Maximum, - - -	2 P.M.	3 P.M.	3.30 P.M.	3 P.M.	3 P.M.

[1] "Exploration of the Air by means of Kites" (at Blue Hill Observatory), *Ann. Astron. Obs. Harv. Coll.*, XLII., Part I., Cambridge, 1897, 100-104; also H. H. Clayton: "Examples of the Diurnal and Cyclonic Changes in Temperature and Relative Humidity at Different Heights in the Free Air," *Blue Hill Met. Obs. Bull.*, No. 2, 1898; H. Hergesell: "Die Temperatur der freien Atmosphäre," *Pet. Mitt.*, XLVI., 1900, No. 5.

Thus, on a well-exposed mountain summit the maximum temperature occurs at about 3 P.M., which is later than in the valleys and earlier than in the free air at the same altitude. In valleys that lie among mountains, the maximum temperature seems to occur about an hour earlier than over neighbouring lowlands.[1]

Stations in valleys at high altitudes have an exceptionally large diurnal range of temperature. In winter, in particular, the greater warming by day under the clearer sky of the higher valleys, together with the correspondingly greater nocturnal radiation and the stagnation of the cold air in the valley bottoms, causes relatively large diurnal ranges of temperature. The maximum diurnal ranges of temperature are probably to be found on dry continental plateaus, like those of Tibet and the interior of Asia as a whole, as well as on the high plains of western North America. Prjewalsky's temperature observations in northern Tibet gave a mean range of 17·3° between 8 A.M. and 1 P.M. even in December; and Säwerzow's observations on the Pamir plateau (3600-4400 m.), in August and September, gave a diurnal range of more than 25° (Woeikof).

While the difference of temperature between 7 A.M. and the afternoon maximum at St. Louis, on the Mississippi River, barely reaches 6·5°, it amounts to 11° at an altitude of 2000 m. in the same latitude on the western plateaus, and the actual daily variation of temperature in these lofty valleys amounts to 16° – 18°. Temperature changes of 25° – 30° within 24 hours are not infrequent; but even here the maximum temperature occurs soon after noon. After what has already been said regarding the simultaneous increase in the intensity of insolation and of nocturnal radiation with increasing altitude, we naturally expect to find great ranges of temperature on plateaus and in high-lying valleys.

[1] J. Hann: "Der tägliche Gang der Temperatur auf dem Obirgipfel und einige Folgerungen aus demselben," *S. W. A.*, CII., 2a, 1893, 709-749; "Beiträge zum täglichen Gang der meteorologischen Elemente in den höheren Luftschichten," *ibid.*, CIII., 2a, 1894, 51-97; "Der tägliche Gang des Barometers an heiteren und trüben Tagen, namentlich auf Berggipfeln," *ibid.*, CIV., 2a, 1895, 505-564; W. Trabert: "Der tägliche Gang der Temperatur und des Sonnenscheins auf dem Sonnblickgipfel," *D. W. A.*, LIX., 1892, 177-250.

In the last named treatise, the author has also attempted to determine how much heat in the air on the Sonnblick is derived directly from solar radiation, and how much is brought up from lower levels by convectional currents. The latter quantity is found to be three times as great as the former.

CHAPTER XVI.

EFFECTS OF MOUNTAINS ON HUMIDITY, CLOUDINESS, AND PRECIPITATION.

Absolute humidity.—The decrease in the vapour contents of the atmosphere with increasing altitude proceeds at a very rapid rate. The decrease in the absolute humidity is indeed much more rapid than the decrease in pressure. The following table shows the relative amounts of water vapour in the atmosphere at several different altitudes. For purposes of comparison, the relative pressure, or the relative density, of the atmosphere at these different altitudes is also included. The vapour pressure and the density of the air at sea-level have the value 1.[1]

DECREASE OF VAPOUR TENSION AND OF PRESSURE WITH INCREASE OF ALTITUDE.

Altitude in meters.	Water Vapour.	Pressure.
0	1·00	1·00
1000	0·73	0·88
2000	0·49	0·78
3000	0·35	0·69
4000	0·24	0·61
5000	0·17	0·54
6000	0·12	0·47
7000	0·08	0·42
8000	0·06	0·37
9000	0·04	0·32

[1] Determined by the formula

$$e_h = e_o\,10^{-\frac{h}{6500}}, \quad \text{or} \quad \log e_h = \log e_o - \frac{h}{6500},$$

in which e_h is the vapour tension at the altitude, h; e_o is the vapour tension at the

These figures are to be interpreted thus: When, for example, the amount of water vapour in the air of central Europe, in summer, is expressed by a vapour pressure of 10 mm., the vapour pressure in the same region, on a mountain 4000 m. high, is only 2·4 mm. On the equator, however, with a vapour pressure of 20 mm. at sea-level, the value at 4000 m. is 4·8 mm. The mean air pressure in both cases is about 470 mm. Half of the total amount of water vapour in the atmosphere is below 2000 m.; about three-quarters are below 4000 m.; and fully nine-tenths are below 6500 m.; while the pressure of the atmosphere is not reduced to one-half of the sea-level pressure until an altitude of 5000-6000 m. is reached.[1]

Mountains, therefore, play an important part with respect to the vapour of the earth's atmosphere. When they rise to considerable altitudes, they may become effective climatic barriers, often separating by sharply defined lines, within a very limited area, districts which are well watered from those which are arid.

Relative humidity, or the degree to which the air is supplied with water vapour, shows no regular change with change of altitude; in fact, as a general rule, it varies but little with such a change. Within the tropics, on mountains where there is abundant rainfall, there is, however, a definite altitude at which the air is almost constantly saturated with water vapour during the rainy season, which may locally extend throughout the greater part of the year. Here there is, consequently, an almost permanent belt of clouds, the altitude of which is usually between 1300 and 1600 m. above sea-level. In higher latitudes, this saturated stratum of air is at a lower altitude in winter, often resting on the ground for days and weeks at a time; while in summer its altitude is much greater. The annual march of relative humidity at considerable altitudes is therefore the opposite of that on the lowlands. On mountains, winter is the drier season, while spring and summer have the highest relative humidity. On the lowlands, on the other hand, winter is the time when the air is usually nearest the point of saturation, or, as we commonly say, when the air is most damp, and summer is the time when the air is farthest from saturation. We have, unfortunately, but few reliable observations of humidity from mountain stations, especially during the winter

lower level, and h must be expressed in meters. See *Z.f.M.*, IX., 1874, 193-200; XIX., 1884, 228-235; *M.Z.*, XI., 1894, 194-197.

[1] See also the discussion by R. Süring in *Wissenschaftliche Luftfahrten*, Vol. III. pp. 157 et sqq.

months, because the psychrometer readings are not to be trusted when the temperatures are below freezing.

ANNUAL MEANS OF RELATIVE HUMIDITY AT DIFFERENT ALTITUDES.

VALAIS ALPS.[1]

Station.	Altitude (meters).	Vapour Tension (mm.).	Relative Humidity (per cent.).
Theodul Pass,	3330	2·6	82
Simplon,	2010	4·1	78
Martigny,	500	6·8	72
CEYLON.[2]			
Newara Eliya,	1875	11·0	83
Kandy,	520	16·9	77
Coast,	—	21·7	79

The annual march of relative humidity at considerable altitudes in the Alps, as compared with that nearer sea-level, may be studied in the following table :—

ANNUAL MARCH OF RELATIVE HUMIDITY (IN PERCENTAGES) AT DIFFERENT ALTITUDES IN CENTRAL EUROPE.

Station,	Theodul Pass.	Sonnblick.	Stelvio.	Säntis.	Vienna.	Geneva.
Altitude (meters),	3330	3100	2470	2467	195	440
Winter,	79*	71*	71*	78*	**81**	**85**
Spring,	**89**	83	**84**	81	77	73
Summer,	80	**86**	78	**84**	64*	70*
Autumn,	83	82	73	83	75	82
Year,	83	80	77	81	72	77

On the high mountains of central Europe the winter is the driest and the clearest season ; while spring and summer are the most moist and

[1] Mean of 1865-66.

[2] Mean of 3 years' simultaneous observations. See also A. Woeikof : "Temperatur und Hydrometeore auf dem Agustia Peak in Südindien und am Fusse des Berges," *M.Z.*, XIII., 1896, 405-416.

the cloudiest. The conditions are just reversed on the lowlands. On Ben Nevis (1343 m.), because of its peculiar situation, the air is almost saturated with water vapour throughout the year, the mean annual relative humidity being 94 per cent., and the means for December and June being 97 and 90 per cent., respectively. On Pike's Peak (4308 m.), the relative humidity is 79 per cent. in winter, 81 per cent. in spring, 75 per cent. in summer, 77 per cent. in autumn, and 78 per cent. for the year. In the tropics, the march of the relative humidity on mountains, as well as near sea-level, follows the rainy seasons.

The daily march of relative humidity at great altitudes is at present known for a few stations only. In the case of the Sonnblick, the observations of humidity for a whole year have been discussed.[1] The results show that the minimum relative humidity occurs between 8 and 9 A.M. throughout the year; while the maximum occurs after noon in winter, and between about 8 and 10 P.M. throughout the remainder of the year. The observations made during the summer on the Faulhorn and at the Grands Mulets, on Mont Blanc, give a similar result, the minimum coming at 10 A.M., and the maximum, between 6 and 8 P.M. On the summit of Ontake (3055 m.), in Japan, the air was also found to be driest about noon, and most damp at 6 P.M. (August).

The daily march of vapour pressure on mountains is characterised by the fact that the amount of water vapour in the air is greatest soon after noon (between 3 and 4 P.M. on the Sonnblick); while it is smallest in the early morning hours. All well-exposed mountain stations, in all climates, agree in showing a rapid increase in the vapour contents of the atmosphere, and also of cloudiness, after noon. The daily march of atmospheric humidity in the valleys, on the other hand, is the same as that on lowlands, but has a somewhat greater range; as is also the case with the daily march of temperature. Because of the fact that valleys are better warmed, the air in them is generally drier than that on well-exposed mountain slopes or summits at the same altitude, although the absolute humidity is the same.[2]

The rapid variations and the great extremes of atmospheric humidity on high mountains are very characteristic features of mountain climates. There are frequent changes from complete

[1] J. Hann: "Die Verhältnisse der Luftfeuchtigkeit auf dem Sonnblickgipfel," *S. W.A.*, CIV., 2a, 1895, 351-401.

[2] For observations of the diurnal changes in relative humidity at different altitudes in the free air, as determined by kite observations at Blue Hill Observatory, see *Annals Harv. Coll. Obsy.*, XLII., 1897, part I., 104-105.

saturation of the air, when clouds rest on the mountain tops, to great dryness. These changes are especially frequent and extreme on high, isolated mountain summits, and they are accompanied by corresponding changes of temperature.[1]

Ascending currents of air bring up water vapour from below, which quickly condenses into clouds. On the other hand, calms and descending air currents cause extreme dryness of the higher strata of the atmosphere. Junghuhn's psychrometer observations on the peaks of the high volcanoes of Java, and his classic descriptions of the succession of weather changes on those mountains, furnish us the most admirable examples of the extraordinary variability in the humidity conditions at these altitudes. On Slamat (3374 m.), the mean relative humidity from June 20 to 22 was 52 per cent.; but it varied from 13 per cent. to 100 per cent. within 24 hours. On Semeru (3740 m.), the humidity on September 26, in the afternoon, was only 26 per cent., with a minimum of 5 per cent. Mats made of pandanus leaves crumbled to dust between the fingers. The face, lips and hands crack open during the dry weather at these heights, and people suffer from excessive thirst.[2]

Similar conditions exist on the summits of the Swiss Alps in fine weather, but the contrast between the humidities aloft and below is not so marked as in the districts near the equator. Martin's observations on the grand plateau of Mont Blanc (3930 m.) gave a mean relative humidity of 38 per cent. for the period, August 28—September 1, 1844; while the relative humidity at Chamonix for the same period was 82 per cent. The minimum on the plateau was 13 per cent.; at Chamonix, 50 per cent. This extremely dry condition of the air alternates with periods of several days of stormy weather during which the air is saturated, and when the mountains above a given altitude are constantly covered with clouds. In the valleys, on the other hand, and on the lowlands in general, the air is seldom saturated during the warmer season, saturation occurring only occasionally during the night and the morning hours. Under these conditions, fogs are formed.

Evaporation.—In connection with the absolute and relative humidity in mountain climates, the amount of evaporation must also be considered. Under similar conditions of relative humidity, temperature and wind velocity, evaporation is much greater on mountains than at

[1] In the treatise referred to in footnote 1, on page 289, there is reference to the spells which occasionally occur on the Sonnblick and on the Obir in winter, when the air is very dry, and there is at the same time an increase in temperature.

[2] Junghuhn: *Java*, Vol. I.

lower levels, because of the diminished pressure aloft. Everything dries much more rapidly at great altitudes. Dead animals are mummified without decaying. Thus we find that meat dried in the air is a national dish even in the lower Engadine, at an elevation of 1400-1600 m. above sea-level. Further, perspiration evaporates rapidly; the skin is dry and parched, and thirst is increased. The relative humidity alone is therefore no sufficient criterion for the evaporating powers of a mountain climate; the diminished pressure makes it possible for the water vapour which has been formed to be distributed much more rapidly through the air, and hence evaporation is accelerated. Evaporation is also aided by the very dry condition of the atmosphere which is occasionally noted during spells of fine weather, as has already been pointed out.

Cloudiness.—The amount of cloudiness on mountains increases in some places with increased altitude, and in other places it decreases. Local conditions are thus the controlling factors in the matter. On the mountains of the tropics, the cloudiness during the rainy season is always greater aloft than below; while these conditions are often reversed during the dry season. In higher latitudes, especially in the Alps, the winter is the least cloudy season on the mountains, while spring and summer are the most cloudy. The annual march of cloudiness on the mountains is thus the reverse of that below. A few illustrations of this point may be given.

ANNUAL DISTRIBUTION OF CLOUDINESS IN INDIA AND CEYLON.

Station.	Altitude (meters).	Mean Cloudiness.	Minimum.	Maximum.
Ceylon.				
Colombo, -	12	5·8	4·1 in February.	7·0 in June and August.
Newara Eliya,	1902	5·5	3·5 ,,	8·0 in June.
Eastern Himalayas.				
Goalpara, -	122	4·6	2·3 in November.	7·6 in June.
Darjeeling, -	2262	6·2	4·4 in December.	8·7 in June and July.
Western Himalayas.				
Roorkee, -	270	3·0	0·8 in November.	6·2 in July and August.
Chakrata, -	2150	4·6	1·6 ,,	8·6 in August.

Large numbers of observations are available in connection with cloudiness in the Alps, and from this abundant material the following means have been derived by the author.

MEAN CLOUDINESS IN THE ALPS.

	Altitude (meters).	Winter.	Spring.	Summer.	Autumn.	Year.
Swiss Foreland, - -	420[1]	7·3	5·8	5·2*	6·2	6·1
Tyrol (Valleys), - -	1300[2]	4·6*	5·8	5·4	5·2	5·2
,, ,, - -	1830[3]	3·7*	4·6	5·0	4·2	4·4
Eastern and Western Alps,	2600[4]	4·6*	6·1	5·6	5·5	5·4
Säntis, - - - -	2467	5·1*	6·1	6·5	6·2	6·0
Sonnblick, - - -	3100	5·2*	7·1	7·3	6·2	6·5

The foregoing data show that the fogs of autumn and of winter are the cause of the higher mean cloudiness of the lowlands as compared with that of the higher valleys and the mountain summits. The latter have a less cloudy autumn and a particularly clear winter. The remarkably clear winter sky in the higher Alpine valleys is one of their most striking climatic advantages; for, together with the dryness of the air and the diminished pressure, it is instrumental in favouring an unusually intense insolation. These climatic peculiarities, combined with the prevailing calms, have given villages like Davos their reputation as winter resorts. The calms, in this case, are due partly to the wind-breaking effect of the mountains, and partly to the absence of local winds, owing to the presence of snow on the ground. The snow covers the whole surface uniformly, and almost does away with differences of temperature. On the other hand, mountains which rise abruptly out of lowlands are more cloudy than the lowlands themselves. Such mountains are often cloud-capped, and thus serve as means of foretelling weather changes. Thus the Schafberg, for example, has a mean cloudiness of more than 6; while the Alpine valleys at the same altitude have a mean cloudiness of barely 5. Pike's Peak, which rises to an altitude of 4300 m. from the dry plateaus of the Rocky Mountains, with their clear skies, has a mean cloudiness of 2·7 in autumn, 3·4 in winter, 4·4 in spring, and 4·0 in summer; the mean for the year being 3·6. In contrast to these conditions, Ben Nevis has

[1] Geneva, Neuchâtel, Zürich, Basel, Altstätten.

[2] Prägraten, Marienberg.

[3] Vent, Sulden, Sils-Maria.

[4] St. Bernard, Theodul, Stelvio, Fleiss, Obir.

a cloudiness of more than 8 in every month but April; the mean for April being 7·9, for January, 8·8, and for the year, 8·4.

The diurnal march of cloudiness on mountains, which has been determined during recent years by means of sunshine recorders, is a very instructive element in mountain meteorology. The following table gives the mean duration of bright sunshine, in hours, on the Obir (2040 m.), and on the Sonnblick (3100 m.). The diurnal march of cloudiness on mountains is very similar in all climates, as will be seen in the discussion of other phenomena which follows. Hence these data are on many accounts of far-reaching importance.

MEAN DURATION OF SUNSHINE ON THE OBIR AND ON THE SONNBLICK (IN HOURS).

Hour.	Obir and Sonnblick.				Year.	Year.
	November, December, January.	February, March, April.	May, June, July.	August, September, October.	Obir and Sonnblick.	Vienna and Klagenfurt.
6—7	0·1	7·0	34·4	17·0	58·5	72·9
7—8	12·8	27·7	42·7	38·9	122·1	104·0
8—9	34·8	37·6	**43·7**	44·2	160·3	136·9
9—10	41·3	41·2	42·4	**45·5**	**170·4**	161·0
10—11	45·0	**41·6**	38·8	44·1	169·5	176·5
11—noon	**46·6**	40·0	31·0	42·4	160·0	182·5
Noon—1	46·0	38·0	29·3	40·8	154·1	**183·7**
1—2	44·6	36·6	30·4	38·3	149·9	182·2
2—3	42·2	33·9	29·4	36·7	142·2	175·2
3—4	33·4	31·8	28·3	34·5	128·0	154·8
4—5	12·3	25·3	27·4	30·4	95·4	113·6
5—6	0·0	11·7	25·5	17·6	54·8	68·3
Forenoon,	180·6	195·1	233·0	232·1	840·8	833·8
Afternoon,	178·5	177·3	170·3	198·3	724·4	877·8
Total,	359·1	372·4	403·3	430·4	1565·2	1711·6

During the winter, the greatest frequency of sunshine on the Alpine summits comes at the noon hours. Every winter month has, on the average, over 15 hours of sunshine from 11 A.M. to noon, and from noon to 1 P.M. In other words, every other day in winter has its full complement of sunshine at noon. Toward spring and summer, the frequency of sunshine about noon continually decreases, both relatively and absolutely. In the early spring, the hours between 9 and 11 A.M.

have the most sunshine. In summer, the greatest frequency of sunshine comes as early as 8-9 A.M. At that season, the sunshine decreases rapidly after 10 A.M., or, in other words, the cloudiness increases rapidly toward noon. Finally, in autumn, the hour between 9 and 10 A.M. has the most sunshine. It is thus apparent that there is a very regular displacement of the sunniest time of day from noon in winter, to 9 A.M. in summer, and then back again through 10-11 A.M. in autumn, to noon in winter. During the summer, not even one day in three has sunshine at noon.

The regularity with which the sunshine decreases at noon as summer approaches may be seen still more clearly when the frequency of sunshine in the three-hour period from 11 A.M. to 1 P.M. in the different months is investigated. We then have the conditions set forth in the following table :—

DURATION OF SUNSHINE ON THE SONNBLICK AND ON THE OBIR FROM 11 A.M. TO 1 P.M. (IN HOURS).

December,	47·0	March, -	39·8	June, -	27·0*	September,	40·4
January, -	46·1	April, -	32·2	July,	36·3	October, -	38·1
February,	44·0	May, -	30·6	August,-	43·7	November,	43·4

The two extremes are found in December and in June. When the days are shortest, *i.e.*, when there is the least insolation, and when the earth's surface is least warmed, the noon hours on mountain tops are the sunniest. On the other hand, these hours are the most cloudy when the days are longest and when the earth's surface is most warmed. These observations show very clearly the effect of the ascending currents of air which result from the warming of the earth's surface. Certain other meteorological phenomena which are related to this same process, either as cause, or as effect, will shortly be considered.

This curious diurnal variation of cloudiness is one of the most characteristic phenomena of mountain climates in all warm countries where ascending air currents are well marked, at least in summer. In higher latitudes, as on Ben Nevis, for example, the diurnal march of sunshine is like that on the lowlands of central Europe, the maximum coming nearly at noon, and the afternoon having more sunshine than the forenoon.[1]

[1] R. C. Mossman : "Sunshine Values at Ben Nevis Observatory," *Journ. Scot. Met. Soc.*, 3rd. Ser., No. IX., 1893, 231-236 (*M.Z.*, X., 1893, 350-352).

In the Alpine lowlands and on the Alpine foreland, the noon hours in winter are not the sunniest, as they are on the Alpine summits, but the noon hours of summer have the maximum frequency of sunshine. The mean duration of sunshine from 11 A.M. to 1 P.M. at Vienna and Klagenfurt, for example, is as follows:—

DURATION OF SUNSHINE, 11 A.M. TO 1 P.M., AT VIENNA AND KLAGENFURT (IN HOURS).

December, -	21·3*	March,	49·7	June, -	52·0	September,	57·0
January, -	32·4	April, -	49·0	July, -	64·2	October, -	38·5
February, -	37·0	May, -	54·0	August,	66·5	November,	23·3

In Vienna and Klagenfurt an average winter month has sunshine during 10 noon hours only, while on the Obir and the Sonnblick it has from 15 to 16 hours. In summer, however, the conditions are reversed. The lower stations have 20 hours of noon sunshine a month, while the upper stations have not quite 12 hours. On the lowlands within the tropics, the forenoon is clearer than the afternoon, which is similar to the condition on the mountains of central Europe.

Influence of mountains on precipitation.—The most important influence which mountains have is in causing condensation of the water vapour of the atmosphere, and thus affecting the frequency and the amount of precipitation. This influence is due to the fact that mountains give rise to ascending air currents, whereby a rapid cooling of the air is brought about, and the water vapour is condensed. Both general and local air movements are affected by mountains for, on the one hand, the prevailing winds are to some extent forced to climb up the slopes of mountain ranges which lie in their path, and, on the other hand, mountains themselves provoke local ascending currents, which will be discussed later. Mountains in all parts of the world thus produce "islands" of more frequent and of heavier rainfall. This action is most distinctly seen in regions in which rain seldom, or never, falls. Thus the high plateaus and the mountains of the central Sahara, Asben and Tibesti, have regular summer rains, and the mountains along the Nubian and Arabian shores of the Red Sea are deluged with thunderstorm rains, while the coast itself either has no rain at all, or is visited only by infrequent showers. In the steppe country of central Asia, wherever there are high mountains, the increased rainfall at a certain altitude favours the growth of trees and forests. The cultivation of the arid lowlands in these districts is entirely dependent upon the

mountain streams which are fed by the melting snow and ice of the higher mountains. Similar conditions are found in the arid portions of western North America. As Loew says :[1]

"In the district lying between the Rocky Mountains of Colorado and the Sierra Nevada of California, and extending south to the Mexican border, it may be stated, as a general rule, that all the country below 1000 m. above sea-level is a real desert, and that between 1000 and 1500 m. is a partial desert. At greater elevations, the ground is more and more covered with vegetation, and at altitudes of 2000 and 2400 m. there are magnificent virgin forests, with rich valleys and numerous springs of water. If this region were all below 1000 m., it would be one vast desert which would have an extent greater than that of half of Europe . . . The Painted Desert and the Gila Desert are separated by the forested mountains of central Arizona. In ascending from the desert across the slowly rising ground to these mountains, the vegetation, which is more abundant at every step, indicates the increasing altitude above sea-level almost as accurately as does the barometer. . . . Above 3500 m. the forests again disappear from the mountains, and the flora becomes more scanty because of the decreasing temperature."

In Arizona, the rainfall varies almost directly with the altitude. The low southwestern section receives less than 125 mm. a year, while the stations on the plateau have from 400 mm. to 500 mm. according to their elevation.[2]

On Kilimanjaro the steppe extends up to 1000 m. Above that height comes the cultivated land, and above that, there are bushes up to 2000 m. The damp virgin forest covers the upper slopes between 2000 and 3000 m., and above the forest come fields of grass, with scattering bushes, up to 4000 m. Similarly on Mt. Kenia, there is steppe below; then the forest-zone begins at 2300 m. and reaches up to 3100 m., while still beyond that altitude comes the alpine zone, reaching up to 4500 m.

Windward and leeward slopes of mountains.—Many mountains have a wet side and a dry side.[3] Such mountains lie more or less

[1] O. Loew : "Die Wüsten Nordamerikas," *Mitt. d. Ver. für Erdkunde in Leipzig*, 1876, 3-13.

[2] A. J. Henry : "Rainfall of the United States," *U.S. Weather Bureau, Bull. D*, 1897, 35.

[3] The climatic differences between the windward and the leeward sides of the central mountains of Germany have been very clearly discussed by Dr. R. Assmann, in *Der Einfluss der Gebirge auf das Klima von Mitteldeutschland*, Stuttgart, 1886.

directly across the path of prevailing winds which carry a large amount of water vapour. In the trade wind belts, the eastern slopes of the mountains are generally the rainy ones, especially when the trade wind blows directly off the ocean. In higher latitudes the western slopes are usually the rainy ones, because outside the tropics the prevailing winds are from the west. In southern Asia, however, where the southwest monsoon is the principal rain-bearing wind, the western slopes have the heaviest rainfall. The wet side of a mountain range is always the side which is toward the principal rain-bringing wind of that particular locality, and this wind is one which blows off an ocean, or from lower latitudes. Mountain ranges whose axes are more nearly parallel to the direction of the prevailing rain-bringing winds have no marked rainy and dry sides. This is the case with the Alps, for example, whose southern and northern slopes both have abundant precipitation.

That the rainy side of a mountain range must be matched by a dry side on the opposite slopes hardly needs any explanation. The side of the mountain facing the moist wind receives all the precipitation which results from the condensation of the water vapour in the ascending current as it passes over the mountain. The amount of precipitation depends on the dew point of the ascending air and upon the temperature to which the air is cooled on its passage across the mountain. If the height of the crest is, say, 2000 m., the air cools about 10° or more throughout its whole mass in crossing this ridge. If, therefore, the lower strata have a temperature of 15°, for example, and if the air is saturated with water vapour, then every cubic meter of air will contain 12·7 grams of water vapour. On reaching the summit and cooling to 5°, the same volume of air can contain only 6·8 grams; or, if we take into account the fact that the volume has increased in the ratio of 76 : 60 = 1·27, it can contain 8·6 grams. In every cubic meter of air under these circumstances, 4·1 grams of water vapour are condensed in the passage over a mountain range 2000 m. high. In the case of an air column 2000 m. high, this condensation will give a rainfall of nearly 8 kilograms per square meter, or 8 mm. of rainfall during the time required by the air to ascend 2000 m. Now, as this process may continue for days, it is easy to conceive of the enormous quantities of rain which can fall on the windward side of such a mountain range. When the air gradually descends to its former level on the leeward side of the mountains, it is warmed during the descent and becomes very dry. The descending air is not saturated with water vapour after its temperature rises above that which prevails at the altitude where condensation occurred. This difference between windward and leeward slopes is still

further accentuated in warmer climates by the fact that while one side is cloudy, rainy, and cool, the opposite, leeward, side has intense insolation under a clear sky, which results in a considerable warming of the lowlands. This further increases the dryness on the leeward side, and is unfavourable to the production of rainfall.

Cloud banners and cloud rings on mountains.—Abbe has made some interesting observations of the effect of mountains upon cloud formation in the free air. He calls attention, in connection with this, to the standing waves formed on the upper surfaces of streams that run over a rocky bottom, which are analogous to the waves formed in the atmosphere. Standing waves of this kind, formed in the free air and capped by clouds, may be seen at the summit of Green Mountain, on the island of Ascension. The clouds form to leeward of the peak and drift away 150 km. under the influence of the steady southeast trade wind. These clouds disappear at night when the air is cooler and drier, and regularly reappear every day.[1]

The cloud rings around mountainous islands in the trade winds will be discussed later. Even low-lying coasts may give rise to clouds, as in the case of Cape Cod and the island of Nantucket, Massachusetts, described by Davis.[2] Cape Cod, which is a long arm of the land, is often marked out in the summer sky during quiet weather by floating cumulus clouds. At the same time, the sky over the mainland to the northwest is covered with large clouds, but the sky over the ocean is clear, except that in quiet weather isolated clouds grow over the sandy island of Nantucket, the island itself not being visible from the mainland.

Examples of great contrasts between the rainfall on the windward and leeward sides of mountains may be found in the following cases. The western coast of Norway has 100-190 cm. of rainfall (latitude 58° – 63° N.), while the interior has 50-60 cm. The outer portion of the great fiord at Bergen has 190 cm.; the central portion has

[1] The so-called "cloud banners" on mountain tops, and the "table-cloth" on Table Mountain, are described in Hann, Hochstetter und Pokorny: *Allgemeine Erdkunde*, 5th edition, Vol. I., Vienna, 1896, 180-183. The interesting cloud formation in the case of the helm wind is described by J. Brunskill and others: "The Helm Wind" (with discussion), *Quart. Journ. Roy. Met. Soc.*, X., 1884, 267-275; "Report of the Committee on the Occurrences of the Helm Wind of Cross Fell, Cumberland," *ibid.*, XI., 1885, 226-238; W. Marriott: "The Helm Wind of August 19, 1885" (with discussion), *ibid.*, XII., 1886, 1-10.

Davis has recently described the occurrence of a helm wind, with the accompanying clouds, in the Cevennes (*M.Z.*, XVI., 1899, 124-125).

[2] W. M. Davis: *Elementary Meteorology*, Boston, 1894, 166.

155 cm.; and the inner portion, only 86 cm. The western coast of Scotland has a rainfall between 120 and 300 cm.; while the eastern coast has only 60-80 cm. The southern island of New Zealand has a rainfall of more than 250 cm. on the western slopes of the New Zealand Alps, while but 60-80 cm. fall on the eastern side of the island. Of tropical and sub-tropical countries, the eastern coast of Australia alone needs mention here. In this case, the coast of New South Wales, between latitude 28·5° and latitude 33·5° S., longitude 153·5° and longitude 151·5° E., has a rainfall of 124-188 cm.; in the interior littoral zone in the same latitudes (longitudes 152·5°–150·5°) the rainfall is fairly uniformly 92 cm., while the amount decreases rapidly inland to less than 30 cm.[1]

On the island of Mauritius during the years 1862-1866, 360 cm. fell on the eastern side at Cluny,[2] while only 71 cm. fell on the west coast, at Gros Cailoux, but 27 km. northwest of Cluny during the same period.[3] On the average, the rainfall on the east coast is twice as great as that on the west coast, at the same height above sea-level.

Effect of mountains on rainfall in the Hawaiian Islands.—The lofty islands of the Hawaiian group, in the northeast trade wind belt, furnish the most beautiful examples of the contrast between a damp, rainy, windward slope and a dry, lee slope, which in some places is almost a desert. On the island of Hawaii itself, Hilo, on the northeast coast, has more than 350 cm. of rain. On the mountain sides, the rainfall increases to nearly 450 cm., and probably even to a very much larger amount. Dutton assumes a rainfall of 600-700 cm. in the region of permanent clouds, at an altitude of 1200-1500 m., and this is not an improbable amount; for, on the windward side of the island of Taviuni, one of the Fiji Islands, in latitude 17° S., within the southeast trade wind, 628 cm. were recorded as the mean of two years at an altitude of 170 m. On the western side of Hawaii, the rainfall at Kailua is but 134 cm., while it is still less on the southern coast. Hilea has but 89 cm., and at that point is the desert of Kau. The most remarkable increase of rainfall from the coast inland toward the mountains is seen on Oahu, near Honolulu. Here 60-90 cm. fall on the coast, but the amount increases with extraordinary rapidity in going up the Nuuanu valley. At a distance of 2 km. north of the city, at 120 m. above sea-

[1] In latitude 33° to 34° S. the mean rainfall is 120 cm. at longitude 151·5° E.; at longitude 150° E., 77 cm.; at 148° E., 57 cm.; at 146° E., 46 cm.; 144° E., 35 cm., and at 142° E., hardly 29 cm. of rain falls.

[2] 14 km. from the S.E. coast, and 305 m. above sea-level.

[3] Gros Cailoux is 3 km. from the west coast; 40 m. above sea-level.

level, 230 cm. fall; and at the summit of the pass, where the altitude is 260 m., the amount is 365 cm. Within a distance of about 8 km., and with a difference of altitude of 250 m., the rainfall increases four-fold, from 85 cm. to 365 cm.[1]

Effect of mountains on rainfall in Java and Burma.—The island of Java also shows locally a very noteworthy increase of rainfall inland from the coast. The following amounts of rainfall are corresponding 16-year means. The stations are on the line of the railroad from Batavia to Buitenzorg.

RAINFALL IN JAVA.

	Stations.					
	Batavia.	Meester Cornelis.	Pasar Mingo.	Depok.	Bodjong Gedeh.	Buiten-zorg.
Distance from Coast (km.),	7	11	17	33	43	58
Altitude (m.), - - -	7	14	35	92	130	265
Rainfall (cm.), - - -	180	180	244	306	373	443

Buitenzorg is south of Batavia, hardly 260 m. higher, but has to the southwest of it the mountain Sálak (2190 m.), and to the southeast the mountain Gédeh, which is 2990 m. high. The ground here begins to rise noticeably.

A very characteristic example of a heavy rainfall resulting from topographic controls, and of a deficiency of rainfall in close proximity to it, is found in Burma. The rainfall on the west coast amounts to 400-500 cm.; while behind the Arakan Hills, in the valley of the Irawadi, the rainfall decreases to below 100 cm., and increases again to the north, with the approach toward the mountains, to 150-200 cm. On the delta, the rainfall is about 250 cm.

Effect of the Alps on rainfall.—The distribution of rainfall in the Alps is also very instructive, although the outer flanks of this mountain system, both to north and south, have abundant rainfall, because the axis of the system is not at right angles to the rain-bringing winds, but is parallel to them. Nevertheless, as the Alpine system is made up of several parallel chains, between which there are longitudinal valleys of considerable length, the effect of the mountains

[1] J. Hann: "Der Regenfall auf den Hawaii-Inseln," *M.Z.*, XII., 1895, 1-14, and "Beiträge zur Kenntniss tropischer Regenverhältnisse," *ibid.*, XVII., 1900, 572-574.

upon the rainfall appears in the fact that these longitudinal valleys have very little rain, while the outer portions of the enclosing mountain chain have abundant precipitation. Thus, the rainfall on the northern side of the northern limestone Alps, at Isny, is 139 cm.; at Tegernsee, 118 cm.; and at Salzburg, 116 cm.; while behind them, in the valley of the Inn, Landeck has 57 cm., and Innsbruck, 87 cm. The northern side of the Bernese Alps, according to Benteli, has 150 cm. of rainfall and more; on the southern side, in the Rhone valley, only 60-90 cm. fall; while still farther south, on the southern slopes of the Pennine Alps, in Piedmont, the amount of rainfall becomes very large, Pallanza having 236 cm., Biella 102 cm., and Ivrea, 139 cm. The examples here given need not be extended further, because many other cases of the same kind may be found in the study of the climatology of special districts. The rainfall of the Inn valley may, however, be referred to here, because it is particularly instructive as illustrating the desiccating effect of high and massive mountain chains upon rain-bearing winds.

MEAN ANNUAL RAINFALL IN THE VALLEY OF THE INN (CM.).

Rosenheim.	Innsbruck.	Landeck.	Remüs.	Zernetz.	Bevers.	Sils.	Castasegna.
138	87	57	57	59	79	95	145

The central portion of the Inn valley, between Landeck and Zernetz, is one of the driest portions of the Alps. In order to appreciate the conditions fully, it must be recalled that the Inn valley has an upper and a lower entrance. The Maloja divide does not close the valley, but the Inn valley descends into the valley of the Maira at Bergell over a hardly perceptible saddle. Thus the rain in the valley of the Inn comes from both the upper and the lower ends. The amount is least in the central portion, especially in the lower Engadine valley, which, in consequence of a change in the direction of the axis of the valley, is shut in on almost all sides by high mountains.

Increase of rainfall with increase of altitude in mountains.—There is an increase of rainfall with increased altitude on mountains, but this increase continues only up to a certain height, and above this there is a decrease. The amount of this increase is not governed by any simple laws. It depends entirely upon local conditions, which may be favourable or unfavourable. The cause of the increase in the amount and the frequency of precipitation with increased altitude above sea-

level is to be found, as has already been explained, in the fact that elevations of land force air currents to rise, and hence to cool. Furthermore, even when there are no general atmospheric movements, local ascending currents occur in mountain regions. Other factors favourable to rainfall among mountains rather than over lowlands are the more abundant covering of vegetation, and the greater dampness of the ground, which together locally increase the amount of water vapour in the air. The ground is therefore less warmed, and there is less radiation of heat from it. These conditions are more favourable to a condensation of water vapour than those over lowlands. Both the amount of precipitation, and especially the frequency, are thereby increased.

Mountains cause an abundant condensation of the water vapour brought by the rain-bearing winds, and then prevent this supply of water from being carried away again by the winds in dry, clear weather. After warm, bright, summer days, the grass, shrubs and trees in the Alpine valleys are covered with drops of dew, which give the appearance of a recent fall of rain. The local ascending air currents carry the water vapour aloft in the afternoon, and the condensation of this vapour gives rise to clouds and thunderstorms, which return the moisture to the earth again as rain. The humidity of the atmosphere, which is thus locally increased through the presence of forests and the abundant supply of water, cannot pass over the surrounding mountain walls without being largely condensed into rain. Mountains thus keep their own moist atmosphere. The abnormal spring rains in the Brahmaputra Valley, in Assam, which begin as early as April during the dry season of northern India, are similarly explained as a result of the presence of extended forests; of an abundant water supply, and of the fact that this district is surrounded by mountains. These various factors unite to increase the humidity of the air; to keep this locally increased humidity confined to a limited area; and to produce ascending currents of air. These rains are due to the condensation of the water vapour which has its source locally, and which cannot be carried away by the winds.

The rainfall increases on approaching a mountain range because the air is forced to rise even at some distance away from the obstruction which lies in its path. Almost all mountainous regions furnish examples of the rule that the rainfall does not increase suddenly, just at the base of the mountains. This has been shown very clearly in the case of the stations on the railroad from Batavia to Buitenzorg, and on the island of Hawaii. The following case, cited by Blanford, shows in

the most striking manner the effect which a considerable elevation has upon the rainfall at some distance away.

INFLUENCE OF PROXIMITY OF MOUNTAINS ON RAINFALL IN INDIA.

Station, - - - - -	Dacca.	Bogra.	Mymensing.	Silhet.
Distance from Khasi Hills (km.)	161	96	48	32
Rainfall - - - (cm.)	191	231	274	380

All these stations are on a level plain, at an elevation of not more than about 20 m. above sea-level. Similarly, even low hills, whose tops are far below the altitude at which clouds form, increase the rainfall for the reason that the upper strata of the air must rise with the lower strata over these elevations. In this ascent, cooling takes place, and if the air is already nearly saturated with water vapour, clouds and rain may result.

The increase of rainfall with increase of altitude may now be illustrated by the following examples:—

INCREASE OF RAINFALL WITH INCREASE OF ALTITUDE IN THE CENTRAL MOUNTAINS OF GERMANY.

Altitude (m.), - -	100-200	200-300	300-400	400-500	500-700	700-1000
Rainfall (cm.), - -	58	65	70	78	85	100

The central mountains of Germany are below the altitude above which the rainfall again begins to decrease. The numerous stations in Bohemia give a mean rainfall of 60 cm. for a mean altitude of 260 m.; of 65 cm. for 450 m.; of 81 cm. for 640 m.; and of 103 cm. for 870 m. The increase with increased altitude is a progressive one up to the height of about 1000 m.

Rawson has shown the increase of rainfall with increased altitude on the basis of observations made at the large number of stations on the island of Barbadoes.[1] This increase is seen in every 30 m. of increase of altitude. The following table gives a general summary of Rawson's results. The successive altitudes, as given in the original paper, are here grouped by threes, and the data are given, in round numbers, in the metric system.

[1] "Report on the Rainfall of Barbadoes," *Barbadoes*, 1874.

INCREASE OF RAINFALL WITH INCREASE OF ALTITUDE IN BARBADOES.

Altitude (meters), - -	50	140	240	over 280
Number of Stations, -	51	25	21	9
Rainfall (cm.), - - -	111	121	142	158

The increase of rainfall with increased altitude on the rainy side of a mountain range, and the rapid decrease on the lee side of the crest-line, are seen in the following summary of mean rainfalls :—

VARIATION OF RAINFALL WITH ALTITUDE ON WINDWARD AND ON LEEWARD SLOPES OF MOUNTAINS.

BLACK FOREST.
(Stations follow in order from West to East.)

Station.	Altitude (meters).	Relative Amounts of Rainfall.
Auggen, - - - - -	290	1·00
Badenweiler, - - - -	420	1·23
Höhenschwand, - - - -	1010	1·76
Donaueschingen, - - -	690	1·01
		Absolute Amounts of Rainfall.
Auggen, - - - - -		107 cm.
Höhenschwand, - - - -		188 ,,

ARLBERG.
(Stations follow in order from West to East.)

		Relative Amounts of Rainfall.
Bludenz, - - - - -	590	1·00
Klösterle, - - - - -	1060	1·15
Stuben, - - - - -	1410	1·44
St. Christoph, - - - -	1800	1·52
St. Anton, - - - - -	1300	0·69
Landeck, - - - - -	800	0·48
		Absolute Amounts of Rainfall.
Bludenz, - - - - -		120 cm.
St. Christoph, - - - -		182 ,,
Landeck, - - - - -		57 ,,

Marriott has recently considered the relation of rainfall in England to altitude above sea-level.[1] The mean annual and mean monthly rainfalls at the English and Welsh stations for the ten-year period 1881-1890 form the basis of the discussion, and the stations are classified according to altitude. The increase of rainfall with altitude is thus summarised:—

30 m. (100 ft.)	+ 9 per cent.	180 m. (600 ft.)	+ 5 per cent.
60 m. (200 ft.)	+ 3 ,,	215 m. (700 ft.)	+ 38 ,,
90 m. (300 ft.)	+ 3 ,,	245 m. (800 ft.)	+ 3 ,,
120 m. (400 ft.)	+ 14 ,,	275 m. (900 ft.)	+ 4 ,,
150 m. (500 ft.)	+ 1 ,,	305 m. (1000 ft.)	− 21 ,,

In California,[2] along the line of the railroad from Sacramento to Summit, the rainfall increases from 49 cm. a year at Sacramento (22 m.) to 115 cm. at Colfax (738 m.), or at the rate of 1 cm. in a little over 10 m. With a further increase of altitude, the rate of increase in rainfall becomes smaller. Thus, at Cisco (1810 m.), the mean annual rainfall is 125 cm., and at Summit (2139 m.), it is 118 cm. Between Colfax and Cisco the rainfall increases at the rate of only 1 cm. in about 100 m., and between Cisco and Summit there is a slight decrease in the mean annual rainfall.

Beyond Summit, the rainfall decreases with extraordinary rapidity toward the east; for, in the valley of the Humboldt River, the annual amount is only 8-10 cm. Further east, the rainfall increases again toward the Wahsatch Mountains.

If a chart be constructed showing the amounts of precipitation by means of lines of equal rainfall (isohyetal lines), we shall have a very striking reproduction of a topographic chart of a country. A beautiful example of this is found in the case of the Bohemian basin, in which the rainfall increases on all sides toward the enclosing mountain walls, and especially in the direction from southwest to northeast.

The altitude of the zone of maximum rainfall, that is, of the zone in which the rainfall is heaviest and above which it decreases again, is a question of great interest. So far, but few observations on this point are available, especially from the tropics, where on account of the greater uniformity of the temperature, the relations are the simplest.[3]

[1] W. Marriott: "Rainfall in the West and East of England in Relation to Altitude above Sea-Level," *Quart. Journ. Roy. Met. Soc.*, XXVI., 1900, 273-278.

[2] Data furnished by Professor A. G. M'Adie, of the U.S. Weather Bureau.

[3] Huber has attempted to show the dependence, in the Canton of Basel, of the annual rainfall upon the altitude, and upon the average angle of slope of the surrounding mountain sides over which the rain falls. The following formula is

Hill found the altitude of the zone of maximum rainfall during the monsoon rains in the northwestern Himalayas to be 960 m. above the plains of the Northwest Provinces, or 1270 m. above sea-level.[1] If the rainfall on the plains is taken as 1, the amount in the zone of maximum rainfall is 3·7 ; and at 3000 m. it is only 0·2. During the period of less rainfall, in winter and spring, when the air is drier, the zone of maximum rainfall is higher. Sykes had previously called attention to the fact that the rainfall of the western Ghats attains a maximum at about 1400 m. above sea-level. In 1849, the relative amounts of rainfall in the Uttra Mullay Range were as follows :—

RELATIVE AMOUNTS OF RAINFALL IN THE UTTRA MULLAY RANGE, INDIA.

	Base,	Attagherry.	Uttra Mullay.	Augusta Peak.
Altitude above sea-level - (m.),	150	670	1370	1890
Relative rainfall (per cent.), - -	1·00	1·72	2·53	1·96

The means derived from several years' observation also show the greatest rainfall at the stations about 1400 m. above sea-level, and the places at which the greatest rainfalls in the world have been recorded, Cherrapunji, in the Khasi Hills (1260 m.), with 1253 cm.,[2] and Mahableshwar (1380 m.), with 643 cm., as well as Baura, with 662 cm., are at this same general height above sea-level. In Java, Junghuhn

obtained, in which h is the relative altitude of the station above 300 m., and a denotes the average angle of slope :

$$\text{Annual rainfall} = 793 \text{ mm.} + 0{\cdot}414\, h + 381{\cdot}6 \text{ tang } a.$$

(R. Huber : *Die Niederschläge im Kanton Basel in ihrer Beziehung zu den orographischen Verhältnissen*, Zürich, 1894).

The stations around the Rigi and the Säntis likewise show an increase of rainfall of 40-50 mm. for every 100 m. of altitude. It appears, nevertheless, from the "topographic" formula, that the influence of the angle of slope is greater than that of the altitude (41 mm. per 100 m., as compared with 382 mm. for a slope of 45°). As the slopes are usually less steep at the crest-line of a mountain range than on the mountain sides, it follows that there must be a zone of maximum rainfall on the sides. (A. Riggenbach : *Verhandl. der naturforsch. Gesells. zu Basel*, X., No. 2.) See also a recent mathematical discussion by F. Pockels : "Zur Theorie der Niederschlagsbildung an Gebirgen," *Ann. der Physik.*, 4te Folge, IV., 1901, 459-480. Translated in *Mo. Weather Review*, XXIX., 1901, 152-159.

[1] S. A. Hill : "Die Höhe der Maximalzone des Regenfalls in Nordwest-Himalaya, und ihre physikalische Bedeutung," *Z.f.M.*, XIV., 1879, 161-165.

[2] 18 to 24 years' observations.

finds the zone of maximum rainfall at an altitude of about 1000 m. There are hardly any observations or studies of the altitude of the zone of maximum rainfall in the case of mountains of the middle and higher latitudes. In the Alps, this zone is probably not far above 2000 m.

In 1868 Prof. John Phillips pointed out (*Rep. British Assoc.*, 1868, p. 472) that the amount of rainfall at Keswick, on the plain at the entrance to Borrowdale (82 m.), was 150 cm., while it increased towards the head of the valley and on the surrounding mountain slopes. Thus at Seathwaite (128 m.) the annual rainfall was 340 cm.; and in ascending Scafell Pike, from Borrowdale, the maximum rainfall of 420 cm. was found at The Stye (330 m.), while on the summit itself (945 m.) the rainfall was only 163 cm. The means of several stations gave the following amounts of rainfall:—West side (average 160 m.), 212 cm.; east side (average 115 m.), 298 cm.[1]; mean, 255 cm. The passes (460 m.) had 335 cm.; the summits (884 m.), 171 cm.; and the maximum rainfall was assumed to occur half-way up Scafell Pike, at an altitude of about 500 m. Mr. G. J. Symons showed (*British Rainfall*, 1898, p. 27) that the more ample data then available did not bear out the earlier conclusions, and that no definite law as to the variations of rainfall with altitude was indicated by the records for the Lake District. Recent observations on both Snowdon and Ben Nevis appear to show that, when the disturbing effect of wind is eliminated, the zone of maximum rainfall is not reached at the highest altitude (1350 m) in Great Britain.

Decrease of rainfall at great altitudes on mountains.—It is easy to see why there must be an upper limit of the zone of maximum rainfall on high mountains, although the same thing is not true of the frequency of precipitation. The vertical decrease of temperature necessarily also involves a decrease in the vapour contents of the air; and the intensity of the precipitation must, consequently, be so much decreased at a certain altitude above sea-level that it can no longer be compensated by greater frequency of precipitation. The maximum rainfall is generally to be looked for at the altitude at which, under the average humidity conditions of the air near sea-level, that air is sufficiently cooled during its ascent to bring about a condensation of the vapour it contains. For, at this altitude, precipitation occurs at the highest

[1] The heavier rainfall of the east side is explained by Laughton as a result of the fact that the air on the lee side of elevations of moderate amount continues to ascend. The rainy stations, Stye Head and Seathwaite, are on the northern side of the mountain. *Quart. Journ. Roy. Met. Soc.*, IV., 80.

temperature of saturation, where the amount of water vapour which is condensed for every degree of temperature decrease is at a maximum.[1]

In winter, when the relative humidity is higher, the altitude of this zone of maximum precipitation is much lower than in summer, and the intensity of the precipitation is less than in the latter season, although the amount may be very considerable by reason of the increased frequency. No investigations have yet been published concerning the seasonal fluctuation of the altitude of this zone of maximum precipitation. The marked relative increase in the amount of precipitation during the winter at the greater altitudes on the mountains of central Germany seems to indicate that these mountains are within the zone of maximum precipitation in winter.

Erk has obtained some results in this connection from two years' observations in Bavaria, on the northern side of the Alps. He finds that there is a seasonal vertical displacement of the zone of maximum precipitation which is chiefly dependent upon the annual march of temperature. A simple zone of maximum precipitation distinctly occurs at altitudes between 600 and 1000 m. above sea-level in winter; but it must be admitted that this zone is not of regular occurrence, nor does it continue throughout the winter.[2]

Zone of maximum rainfall and human settlements.—Sewerzow's interesting notes on the Tianschan, in central Asia, afford a very pretty illustration of the change of altitude of the cloudy and the rainy belt between winter and summer. The zone of the clouds from which the winter snows fall is here found at an altitude of 2500-3000 m., and this is also the altitude at which the evergreen forests occur, these being absent lower down because of the aridity. The higher slopes receive but little snow in winter, but they do have abundant rains which fall from the higher clouds during the warmer months. These rains are favourable to the growth of grass at these altitudes, and hence good pasturage is found there. Thus it happens that the Kirghizes have their winter camps at these great heights, which are almost free from snow, and furthermore offer excellent fodder for their

[1] Temperature,	−10°	−5°	0°	5°	10°	15°	20°	25°	30°
Water Vapour per cubic meter (grams),	2·28	3·38	4·87	6·79	9·36	12·74	17·15	22·83	30·08
Condensation for each 1° of Temperature Decrease (mm.),	0·17	0·25	0·33	0·43	0·57	0·75	0·98	1·25	1·59

[2] F. Erk: "Die Vertikale Vertheilung und die Maximalzone des Niederschlags am Nordabhange der bayerischen Alpen im Zeitraum November 1883 bis November 1885," *M.Z.*, IV., 1887, 55-69.

flocks and herds. In the winter, the Bogintses drive their herds of horses onto the plateau between Barskoun and Narin, at an altitude of 3400-3700 m. above sea-level, among the Sary-Tur mountains. Here there are valleys which are almost wholly free from snow, and hills which supply good fodder, and whose sunny slopes are sheltered from the wind. At the point where the Amu Daria leaves Victoria Lake, in the Pamir, Wood found winter camps of the Kara-Kirghizes at an altitude of 4880 m., and there they pastured their horses, sheep and yaks. The plateau was free from snow in January, and the pastures were open, while the road up to the plateau was deeply covered with snow. These lofty pastures, which have no snow in winter because they are higher than the winter snow clouds, and at the same time are below the snow line, constitute a notable peculiarity of the lofty mountain regions of central Asia. The bright winter sky of the high valleys in the Alps, at 1300-1800 m. above sea-level, offers but a poor meteorological analogy to these conditions.

CHAPTER XVII.

SNOW-LINE AND GLACIERS: CLIMATIC ZONES ON MOUNTAINS.

Snow-line.—The altitude to which the continuous snow cover of high mountains retreats in summer is known as the *climatic snow-line*, and this essentially coincides with the lower névé limit. The snow-line is chiefly controlled by the depth of the winter snowfall, and by the temperature of the summer. Where on a fairly level surface, under conditions of normal exposure, the snow cover is able to maintain itself through summer, there is the boundary of the "eternal snow."

The first stage in the development of the conception of the snow-line, from Bouguer to de Saussure, was dominated by the idea that snow-line and frost-line (*i.e.*, the mean annual temperature of freezing) are identical. This purely physical view led to the framing of a general hypothesis as to the course of the snow-line, and also to the development of a formula which expressed the height of the snow-line simply as a function of the latitude.

In the second stage, under the lead of de Saussure, the conception of the snow-line began to be distinguished from that of the frost-line, after it had been recognised that these two lines by no means actually coincide. Observation of the actual conditions in nature began to be regarded as of greater importance.

Under the leadership of Alexander von Humboldt, the comparative geographic method of looking at the subject totally displaced the deductive method. The real complexity of the problem was recognised, and the idea of the climatic snow-line as the lower limit of the permanently continuous snow-cover in mountains began to be formed. The view which prevails at the present day may be considered as the last stage in the development of the conception which recognises

an orographic snow-line, as well as a climatic snow-line.[1] The credit of establishing and enforcing the conception of the orographic snow-line, together with that of the climatic snow-line, belongs to Ratzel. This orographic snow-line is the lower limit of snow fields and of névé patches, which occur as isolated patches, or in considerable numbers, and which owe their permanent preservation essentially to favourable orographic surroundings.[2]

Bouguer believed that the climatic snow-line coincides with the isothermal surface of 0°. Humboldt and Buch substituted for the mean annual isothermal surface of freezing a mean summer temperature of 0°. Renou sought to prove that the snow-line is to be found in all climates at the altitude at which the mean temperature of the warmest half of the year is zero.

The amount of precipitation, and especially the amount of precipitation which falls during the winter months as snowfall, is a factor which, in addition to temperature, plays so important a part in this connection that, as was pointed out by von Humboldt, many phenomena would be inexplicable unless the effect of precipitation were taken into account. Furthermore, among the local influences which also play a part may be noted the exposure of the mountain sides to the sun, and to warm and dry winds from the land; the steepness of the slopes, and the height to which the mountains rise above the region of snowfall. The fact that the altitude of the snow-line is lower on the northern than on the southern slopes of mountains in the northern hemisphere, is explained by the more intense insolation on the latter.

Height of the snow-line: snowfall and snow-line.—Near the equator, the snow-line has a well-defined lower limit, and is everywhere at about the same height above sea-level. The case is quite different in higher latitudes, where the exposure, the steepness of the slopes, and other local influences play a very important part. The course of the snow-line consequently becomes very irregular, and its mean height above sea-level is rather difficult to determine. In higher latitudes, the snow-line has by no means the same altitude throughout the same mountain range, and isolated fields and patches of snow occur in shady ravines and gullies far below the true snow-line (Ratzel's orographic

[1] F. Klengel: "Die Historische Entwickelung des Begriffs der Schneegrenze von Bouguer bis zu A. von Humboldt," 1736-1820, *Mitth. d. Ver. für Erdkunde in Lepzig*, 1888, 105-190.

[2] F. Ratzel: *Zur Kritik der sogenannten Schneegrenze*, Leopoldina, 1886. Compare also F. Ratzel: "Die Schneedecke besonders in deutschen Gebirgen," *Forsch. zur deutsch. Landes u. Volkskunde*, IV., No. 3, Stuttgart, 1889.

snow-line). While on the equator the lower limit of snow remains at about the same height above sea-level throughout the year, the fluctuations which this line undergoes increase with the latitude, and with the accompanying increase in annual range of temperature. At some places within the tropics, as, for example, in Mexico, the lower limit of snow moves nearer sea-level during the rainy season, which corresponds to our summer, and moves farther up the mountains during the dry and sunny, although cooler, season of winter. The lower limit of individual snowfalls during the summer rainy season may be fixed at 3600 m. in the Andes at Quito; while the snow-line itself is 4500 m. above sea-level. On the southern side of the Himalayas, it snows almost every winter down to 1500 m. above sea-level, and, in very rare cases, snow has fallen even at 900 m. above sea-level. H. Schlagintweit gives the following data for the height of the snow-line in the Himalaya Mountains, in each of the four seasons:—

HEIGHT OF THE SNOW-LINE IN THE HIMALAYA MOUNTAINS (In meters).[1]

	Winter.	Spring.	Summer.	Autumn.
Northern Slope, - -	2700	3800	4900	4270
Southern Slope, - -	2600	4270	5200	4700

According to Drew, the plain of the Punjab, about 300 m. above sea-level, is free from snow, as are the outer mountain ranges. Snow may fall at an altitude of 1200 m. in January, and at 3000 m. it remains on the ground for about three months. The valleys hold the snow longer than this, the whole of Ladakh being snow-covered for more than three months. Yet, even at a height of 4000 m. above sea-level, the thin cover of snow disappears under the influence of evaporation and of the wind. The Champos pasture their cattle during the winter at an altitude of 4200 m. in the broad alluvial valley of the upper Indus. At Leh, the summer snow-line is at an altitude of 5600 m. on the northern side and of 5800 m. on the southern side; while, in the eastern part of Rupshu and around Pangkong, the height is 6100 m.

According to Hill, the lower limit of snowfall in the Northwest Provinces of India is 1700 m. in winter in the outer ranges of Kumaon, and somewhat lower in Dehra Dun and in the northern Punjab. About once in ten years snow falls as low as 1500 m., and the lowest point at which snow has fallen in the first half of the nineteenth century was 900 m. above sea-level.

On the island of Teneriffe (latitude 28° N.), the lowest snowfall comes at a height of about 1300 m. above sea-level; on the island of Madeira

[1] The altitude at which the snow-cover remains for at least half the year.

(latitude 32° N.), the limit is 800 m., and at latitudes 36-37° N., in Algeria and in southern Spain, snowfalls occasionally take place at sea-level. The upper limit of the snow-line is above 3000 m. in the Sierra Nevada Mountains.

Equatorial limit of snowfall.—Humboldt summarises as follows the results of his observations concerning the lower limit of snowfall :—

ALTITUDE OF SNOWFALL AND OF SNOW-LINE (METERS).

Latitude.	Lower Limit of Snowfall.	Lower Limit of Eternal Snow.	Difference.
0°	4000	4800	800
20	3000	4600	1600
40	0	3000	3000

These figures are only approximate, for it occasionally, although infrequently, snows at Naples, Lisbon and even at Malaga, and snow has been recorded in the city of Mexico, in latitude 19° N., and 2270 m. above sea-level. In the latter case, however, no snowfall had previously been recorded for hundreds of years. On this same occasion the streets at Valladolid (in Mexico; latitude 19° 42′ N., altitude 1950 m.) were covered with snow for several hours.

Although the limits of winter snowfall at sea-level may be considered to be about latitude 36° N. in southern Europe, and although the district of the Mediterranean Sea, as a whole, lies within the equatorial limit of snowfall, in eastern Asia it occasionally snows as far south as Canton (latitude 23° 12′ N.), on the Tropic of Cancer. This, however, is the extreme limit. Next come the southern United States of America, where snow occasionally falls on the lowlands as far as latitude 26° N. In regard to the snowfall of the Southern Hemisphere, it may be said that snow has fallen once at Sydney (latitude 33·9° S.), New South Wales; and it has snowed at Buenos Aires and at Montevideo, in South America. Table Mountain, at the Cape of Good Hope, is, however, rarely snow-covered.

The equatorial limits of snowfall, according to the careful investigations of Fischer, are as follows [1] :—

[1] Hans Fischer: "Die Aequatorialgrenze des Schneefalls," *Mitt. d. Ver. für Erdk. zu Leipzig*, 1887, 99-274, with chart.

EQUATORIAL LIMITS OF SNOWFALL.

	Regular Snowfall.	Occasional Snowfall.
	°	°
Europe, west coast,	45 N.	33 N.
Coasts of Mediterranean Sea,	37 N.	29 N.
Asia, interior,	24 N.	22 N.
,, east coast,	30 N.	22·5 N.
North America, west coast,	45 N.	34 N.
,, ,, interior,	30 N.	19 N.
,, ,, east coast,	35 N.	29 N.
South Africa, interior,	—	24 S.
Australia, east and south coast,	—	34 S.
South America, west coast,	45 S.	34 S.
,, ,, east coast,	44 S.	23 S. (?)

Snow falls occasionally throughout all southern Europe; in Tripoli, Algiers, lower Egypt, and over the whole of Syria and Mesopotamia. At the Cape of Good Hope, snow falls only occasionally in the interior. On the plains of the interior of South America, snow falls nearly to the tropics.

Seasonal variation in the height of the snow-line.—A 30-year series of daily observations of the height of the snow-line on the Säntis (2500 m.), and in northeastern Switzerland as a whole, has been discussed by Denzler. According to this authority, the number of days during which the snow-cover remains at different altitudes is shown in the following table :—

DURATION OF SNOW-COVER AT DIFFERENT ALTITUDES ON THE SÄNTIS.

Altitude (Meters).	Number of Days.
650	77
1300	200
1950	245

Thus it appears that at an altitude of 1950 m. there are, on the average, only 120 days on which there is no snow on the ground (latitude 47° N.). The retreat of the lower snow-line in spring, and

the descent in autumn, are shown in the accompanying table, the numbers 1, 2, 3 indicating successive decades of each month.

ALTITUDE OF THE SNOW-LINE ON THE SÄNTIS IN DIFFERENT MONTHS (METERS).

Decade.	March.	April.	May.	June.	July.[1]	October.	Nov.	Dec.
1	690	810	1220	1750	2340	1980	1190	820
2	730	900	1250	1930	Sept.	1730	1000	740
3	730	1020	1470	2060	2030	1510	870	—

The snow-line in eastern Switzerland seems to reach its least altitude at the end of January, and its greatest about August 10. It retreats slowly to the upper slopes in spring, and rapidly descends again in the autumn. In the middle of March the altitude of the snow-line is the same as in the middle of December; and at the end of October it is greater than at the end of May. This corresponds to the slow advent of spring at the greater altitudes on mountains, and to the extension of the summer well into the autumn. A second extended series of observations (1863-1878) of the seasonal variation in the height of the snow-line, in this case for the northern Alps at Innsbruck, was carried on by Anton von Kerner, and discussed by Fritz von Kerner.[2]

MEAN HEIGHT OF THE SNOW-LINE IN THE NORTHERN TYROL (METERS).

SOUTHERN EXPOSURE (Northern Slopes in the Inn Valley).

Dec.	Jan.	Feb.	Mar.	April.	May.	June.	July.	Aug.	Sept.	Oct.	Nov.
740	650	740	960	1270	1700	2190	2680	3130	3210	2150	1300
NORTHERN EXPOSURE (Southern Slopes in the Inn Valley).											
680	590	600	720	1100	1540	2030	2470	2930	2760	1890	1010

1 The first decade of July and the last decade of September are alone included, because the snow-line was often above the summit of the Säntis in midsummer, so that the mean altitudes for this period were too low. Thus the highest mean for the second decade of August is 2460 m., while the snow-line is at 2600 m. For a similar reason, the winter observations are omitted.

2 F. von Kerner: "Untersuchungen über die Schneegrenze im Gebiete des mittleren Innthales," *D. W. A.*, LIV., 1887, Pt. II., 1-61 [*M. Z.*, V., 1888 (30)-(33).

The increase in height is most rapid from May to June (490 m.); the decrease is most rapid from September to October (about 900 m.).

The following table gives the duration of the snow-cover with a northern exposure, and also the number of days with snowfall:—

DURATION OF SNOW-COVER AND DAYS WITH SNOWFALL IN THE NORTHERN TYROL.

Altitude (meters), -	600	800	1000	1200	1400	1600	1800	2000	2200	2400
Snow-cover (days),	86	102	122	134	163	194	214	231	253	285
Snowfall (days), -	—	—	142	170	191	209	227	246	267	285

The isotherm of 0° is below the snow-line from December to February; in January, it is 700 m. below the snow-line. During the remainder of the year the isotherm of 0° is above the snow-line, the distance being 1160 m. in June and July.

Temperatures at the temporary snow-line.—The mean monthly temperatures corresponding to the altitudes of the temporary snow-line in the northern Alps, as determined by Denzler and Kerner, are as follows (latitude 47° N.):—

MONTHLY TEMPERATURES AT THE TEMPORARY SNOW-LINE IN THE NORTHERN ALPS.

Height of Snow-line (in hectameters).

March.	April.	May.	June.	July.	August.	Sept.	Oct.	Nov.	Dec.
7·1	10·2	14·4	19·3	24·8	28·6	25·6	18·0	10·0	7·0
Mean Temperatures at the Snow-line.									
2·3°	5·7°	6·7°	7·3°	6·2°	4·0°	3·3°	2·9°	0·4°	−2·3°

In May and June, the mean temperature at the temporary snow-line is thus seen to be 7°. The heavier the snowfall, the higher must be the temperature at the temporary snow-line in the spring. According to Eller, in the case of St. Gertrud, in the Sulden valley, on the Ortler (latitude 46·5° N. altitude 1840 m.), the country thereabouts is free from snow in the middle of May, when the mean temperature is 5°. Yet when the winter snowfall has been heavy, the snow has been found to stay on the ground until June 1, when the temperature was as

high as 6·5°. The pass over the Arlberg (1790 m.) is free from snow on May 20, when the mean temperature is 7°. This corresponds very closely with the results of observation on the Säntis and at Innsbruck.[1]

Hertzer has made some very careful observations, extending over many years, of the height of the temporary snow-line in the Harz Mountains.[2] The following table shows the number of days during which snow lies on the ground at different altitudes in the Harz Mountains, the results being based on 32 years' observations :—

DURATION OF SNOW-COVER IN THE HARZ MOUNTAINS.

Altitude (meters),	240	400	550	700	850	1000	1150
From - - -	Dec. 27	Dec. 14	Dec. 6	Nov. 28	Nov. 21	Nov. 15	Nov. 9
To - - -	Feb. 24	Mar. 5	Mar. 19	Mar. 29	Apr. 5	Apr. 25	May 13
Days, - - -	60	82	104	122	136	162	186

The mean temperature at the temporary snow-line is about 1·5° in the middle of March; 3·2° in the middle of April; and 5.5° in the middle of May, which corresponds to the St. Gertrud case. The summit of the Brocken has no permanent snow-cover for about half the year. At the end of November and at the beginning of April, the snow-line is at about the same height in the Harz and on the Säntis, namely, 900 m.

To the foregoing observations we may add a few data which have been obtained by Birkner regarding the duration of the snow-cover in the Erzgebirge of Saxony.[3]

DURATION OF SNOW-COVER ON THE SAXON SLOPES OF THE ERZGEBIRGE.

Approximate Altitude (meters),	150	250	350	450	550	650	750	880
Duration of Snow-cover (days),	55·4	67·6	80·2	86·2	96·0	117·7	145·4	150·5

The duration of the snow-cover does not increase regularly with increased altitude, but the general average is an increase of 13 days

[1] J. Hann : *Wärmevertheilung in den Ostalpen.*, Zeitschr. d. deutsch. u. oesterr. Alpenver., 1886, 48, et sqq.

[2] *Schrift. d. naturw. Ver. des. Harzes*, I., 1886.

[3] O. Birkner : "Die Dauer der Schneedecke im Bereiche des sächsischen Erzgebirges," *M.Z.*, VIII., 1890, 201-205.

per 100 m. The following data are taken from another treatise by the same writer. They are typical of the effect of altitude upon the dates of first and last snowfall, frost, and the total annual precipitation.

DATES OF FIRST AND LAST SNOWFALL AND FROST ON THE NORTHERN SLOPE OF THE ERZGEBIRGE.

Altitude (meters), -	100	300	500	700	900 (summit)
First Snow, - -	Nov. 9	Oct. 30	Oct. 27	Oct. 19	Oct. 10
Last Snow, - -	Apr. 18	May 1	May 10	May 17	June 1
First Frost, - -	Oct. 13	Oct. 3	Oct. 1	Sept. 19	Sept. 21
Last Frost, - -	Apr. 28	May 6	May 15	May 31	May 24
Precipitation (mm.),	580	700	800	880	990

The dates of first and last frost at 900 m. (Sept. 21 and May 24) show the influence of the southern slopes, which extends to the crest of the mountains.

The height of the snow-line in different mountains cannot be considered in detail in this volume. For complete data, the reader should refer to Heim's *Handbuch der Gletscherkunde*, pp. 18-21; and to the tables prepared by Berghaus, and published in Behm's *Geographisches Jahrbuch*, Vol. I., 1866, pp. 256-271, and Vol. V., 1874, pp. 472-485. We can here refer only to those facts which bring out most clearly the combined influence of temperature and altitude upon the snow-line.

Near the equator, the snow-line is found at a height of 4000-5000 m., and here the controlling influence of precipitation is shown with great distinctness. Thus, in the Andes of Quito, the snow-line is at an average height of 4560 m. above sea-level on the eastern Cordilleras, which have the heaviest rainfall; while the altitude averages 4740 m. on the drier, western Cordilleras. On Kilimanjaro (lat. $3\frac{1}{4}°$ S.), the snow-line is found at about 4600 m. above sea-level on the moist southern and western slopes, while the altitude is 5500 m. on the dry northern and eastern slopes. On Kenia (lat. $\frac{1}{4}°$ S.) and Ruwenzori (lat. $\frac{1}{2}°$ N.), the snow-line seems to have about the same height, viz., 4500 m.

The snow-line by no means decreases in altitude regularly with increasing distance from the equator. On the contrary, in the dry sub-tropical zones of both hemispheres it locally reaches the greatest altitudes at which it is ever found; viz., over 5000-6000 m. In central Peru, the snow-line in the eastern Cordilleras is at 4870 m., while it is at 5230 m. in the drier western Cordilleras. In southern Peru, with increasing dryness, the snow-line retreats to nearly 6000 m. In Chile,

the height of the snow-line at latitude 30° S., is 4900 m. (Pissis); at latitude 32°-33° S., it is 4200 m. (Güssfeld) and at latitude 34°-35° S. it is between 3100 and 3500 m. South of latitude 37° S., where the more rainy, and finally the very rainy zones, are entered, the snow-line rapidly descends to 1800-2100 m. On Mt. Villarica (lat. 39·5° S.), the height of the snow-line is but 1600 to 1700 m.; on Osorno (lat. 41° S.), the height is from 1400 to 1500 m.; on Corcovado, it is 1360 m.; and at the Strait of Magellan, in a latitude corresponding to that of northern Germany, the height is but 1000 m. The low summer temperatures and the heavy precipitation on the west coast of South America, south of latitude 40° S., give rise to a very curious phenomenon. The snow-line there almost coincides with the tree-line, as was observed by Pöppig, and later confirmed by Philippi. Elsewhere there is always an intermediate zone, as is illustrated by the high pastures of the Alps, which have a vertical extent of about 800 meters.

In North America, at latitude 19° N., in Mexico, the snow-line is at 4400-4800 m., but it does not appear at all in the dry sub-tropical latitudes; and at altitudes of 4000-4500 m. there are only isolated patches of snow ("orographic snow-line"). As soon, however, as the precipitation increases, at about latitude 40° N., the snow-line descends. On Mt. Shasta, the height of the snow-line is 2400 m.; in the Cascade Mountains, at the northern boundary of the United States, it is 2000 m.; on Vancouver Island, it is between 1600 and 1800 m.; and on Mt. St. Elias, at latitude 60° N., it is 800 m.

The position of the snow-line in the Himalaya Mountains is especially instructive. Between latitudes 27° and 34° N., on the Indian side of the mountains, which is the rainy side, the average height of the snow-line is 4900 m.; while on the northern side, toward Tibet, it is 5600 m. In the Karakoram and Kuenluen Mountains (lat. 35°-36° N.), the height reaches 5500-6000 m. The reasons for this increase in altitude of the snow-line on the more northerly ranges are to be found chiefly in the increasing dryness of the atmosphere; in the scanty precipitation, and, to a less degree, in the excessive heating of the plateaus during the summer.

Hooker, Strachey, the brothers Schlagintweit and others have shown that, notwithstanding the higher temperature, the snow-line is lower on the southern side of the Himalaya Mountains, in consequence of the heavier precipitation, than it is on the northern side, which borders on the colder plateau of Tibet, where the rainfall is very much less. This condition, which has been verified by the statements of every observer, has been found by Diener to exist also in the central Himalayas. It must, however, by no means be supposed that these differences in the height of the snow-line will appear on the mountain ridge

which forms the actual divide; or on any individual range of the great mountain complex. In every individual range the snow-line is actually lower on the northern slopes than on the southern. This normal difference between the northern and southern slopes of each individual mountain range is, however, more than balanced by the abnormal rise in the snow-line on the mountain range which follows next on the north. Even in the case of this latter range, considered by itself, the snow-line is lower on the northern side than on the southern; but it is considerably higher on the southern side of this range than on the corresponding side of the next adjoining range on the south. Thus, when the mountain mass is looked at as a whole, the snow-line on the Tibetan side is at a considerably greater altitude than on the Indian side.[1]

In the Pyrenees there is no snow in midsummer on the dry southern slopes, which face the hot plateaus of the Iberian peninsula, but on the northern slopes the snow-line may be found at a height of 2800 to 2900 m.

In the Caucasus, differences similar to those in the Himalayas are found; and these differences are caused by corresponding differences in precipitation. The following table illustrates the conditions in the Caucasus:—

HEIGHT OF SNOW-LINE IN THE CAUCASUS MOUNTAINS (METERS).

Principal Range.	Latitude.	Mean Height of Snow-Line.	
		Northern Slopes.	Southern Slopes.
Western Portion,	42°·7 – 43°·5 N.	3400	2920
Central Portion,	41·5 – 42·7 N.	3300	3230
Eastern Portion,	40·5 – 41·5 N.	3600	3720

In the western portion of the Caucasus, the snow-line is at a less altitude on the southern slopes, because the precipitation is much heavier there. This difference in height decreases to the east, and the general increase in dryness, and in the summer temperature, brings about a considerable increase in the height of the snow-line.

In the Alps, there is an increase in the height of the snow-line toward the central portion of the chain; where there is less precipitation, and where a rise in the isothermal surfaces accompanies the general increase in elevation of the surface. In the western Alps, the

[1] Carl Diener; "Schneegrenze und Gletscher im Zentralhimalaya," *Deutsche Rundschau für Geographie*, XVI.

snow-line is between 2700 and 2800 m. above sea-level;[1] in the Bernina and Ortler group it rises to over 2900 m.; in the high Tauern it is 2600 m. on the northern side, and 2800 m. on the southern.[2]

The northern ranges of the eastern Alps, as well as the southeastern ranges (Julian Alps), have a low snow-line because of the heavy rainfall.

The increase in the height of the snow-line in Norway, from the coast toward the interior, was demonstrated by the classical investigations of Wahlenberg and Buch. In latitudes 70-71° N. the snow-line at the coast is found at 700-800 m. above sea-level; while the height is 1000 m. in the interior. At latitude 62° N., the height at the coast (Aalfotbrae) is 1200 m.; somewhat farther inland, on Folgefond (lat. 60° N.), it is 1450 m.; on Jostedalsbrae (lat. 61° N), it is 1600 m.; and at Jotunheim (lat. 61° N.), it is between 1800 and 1900 m.[3] In the same latitude, the snow-line is found about 400 m. nearer sea-level on the west coast of North America, which has a heavier rainfall and a much lower temperature.

In Iceland, the mean of twelve observations by Thoroddsen fixes the height of the snow-line, in latitudes 64-65° N., at 870 m. above sea-level. It is lower on the southern side (600 m.) than on the northern side (1300 m.), because of the heavier precipitation on the coast. On Myrdalsjökul, the height is 900 m. on the south, and 1150 m. on the north, the mean being about 1000 m.

In Spitzbergen, in latitude 77° N., in the Hornsund, the height of the snow-line is 460 m. In Franz Joseph Land, in latitude 82° N., it is 100-300 m. In the regions about the north pole, the snow-line has nowhere been found to descend to sea-level. In the high latitudes of the Antarctic, on the other hand, it reaches to the level of the sea. The winter is relatively mild there, but the summer is cold. On the island of South Georgia, in latitude 54·5° S., Vogel believes that the snow-line may be fixed at 550 m. above sea-level.

The mean annual and mean summer temperatures at the snow-line (*i.e.*, at the limit of eternal snow) vary very greatly under different climatic conditions. The heavier the precipitation and the smaller the annual range of temperature, the higher is the mean temperature at the snow-line. In the Andes, at Quito, on the equator, the temperature at

[1] The effect of exposure in the district of the Dammastock is shown by R. Zeller to be as follows: Northern slopes, 2740 m.; eastern, 2780 m.; southern, 2870 m.; western, 2860 m.

[2] Compare the instructive chart given by Richter in *Die Gletscher der Ostalpen*, Plate 4.

[3] S. Richter: "Die Gletscher Norwegens," *Hettner's Zeitschr.*, 1895-96.

the snow-line in the eastern Cordillera, is + 3°; in the western Cordillera, + 2°. In southern Chile, also, the mean annual temperature at the snow-line is about 3°. On the southern slopes of the Himalayas, from Sikkim to the Northwest Provinces, the snow-line coincides with the isotherms of + 0·5° to − 1°, and with a July temperature of 6·7°. On the Tibetan slopes of the Himalayas, on the other hand, a mean temperature of − 4° to − 5° prevails at the snow-line. In the eastern Alps, a mean annual temperature of about − 3°, and a summer temperature of 3° to 4°, are found at the snow-line. On Nova Zembla and Spitzbergen, a mean annual temperature of − 10° to − 11° corresponds to the height at which perpetual snow is found; and in the interior of Asia, in northern Siberia, the mountains are not snow-covered although the mean temperature is − 17°. On the coast of northern Siberia, Nordenskjöld found mountains 600 m. high without any snow in summer. An extreme continental climate, with a slight precipitation, makes a permanent snow-cover impossible.

Lower limits of glaciers.—The altitude above sea-level, and the mean annual temperature of the regions down to which the lower ends of glaciers descend, depend much more upon the local conditions than is the case with the snow-line. The position of the foot of a glacier also depends upon the extent of the ice-field from which the glacier is supplied, and upon the slope of the bed, *i.e.*, upon the amount of forward pressure from up-stream, and the rapidity of movement of the ice down-stream. The sooner and the more completely the losses by melting are made up, the higher will be the temperature of the regions into which the glacier may descend. Where there is little precipitation, and where the summer is hot, the glaciers must end where the mean temperatures are still low. A few examples will serve to illustrate these different conditions.

West coast of New Zealand: The Franz Joseph glacier (lat. 43° 35′ S.) ends at 290 m. above sea-level, and the Fox glacier, at 200 m. The mean annual temperature is 10°, which is higher than that of Vienna. On the eastern side of the mountains, in the same latitude, the great Tasmanian glacier descends only to 780 m. above sea-level.

West coast of Chile: A glacier in latitude 46·5° S. reaches sea-level. The mean annual temperature is 8·4°.

Northwest coast of North America: A glacier at the end of a fiord east of Fort Simpson, in British Columbia (lat. 54° N.), descends into the sea. The mean annual temperature here is 10°, according to Dall.

Karakoram Mountains: The Biafo glacier (Balti), in latitude 35° 41′ N., ends 3080 m. above sea-level. The temperature is about 9°.

Himalayas : The Chaia glacier (Gharval; lat. 31° N.) stops at 3200 m. above sea-level, with a mean annual temperature of 7°.

Alps : The mean height above sea-level of the lower ends of eight primary glaciers on Mont Blanc is 1450 m. The mean annual temperature is about 4·2°. The Bossons Glacier descends as far as the isotherm of 6·5°.

In the mountains about Oetzthal, where the climate is more continental in character, and where there is less precipitation, the lower ends of the ten largest primary glaciers are 2100 m. above sea-level, according to Sonklar; the mean annual temperature at that point being – 0·1°, and the summer temperature, 7·8°. In the Altai Mountains of western Siberia (lat. 50° N.), the Katun glacier descends to 1240 m. above sea-level, where the probable mean annual temperature is – 1·7°; and in eastern Siberia, on Munku-Sardyk (lat. 52° N.), a glacier on the southern side reaches down to 3170 m. above sea-level, where the mean annual temperature is about – 10°.

Thus the lower ends of glaciers are seen to lie in districts whose mean annual temperatures differ at least 20° from one another. The influence of climate upon the development of glaciers is shown with remarkable clearness in Alaska. Along the coast, where there is abundant precipitation, and where the summer temperatures are very low, the glaciers are both numerous and large. In the district about Mt. St. Elias there are hundreds of great rivers of ice which descend to sea-level, some of them even reaching into the ocean and breaking off there, with ice walls 100 m. high. In the interior, however, in the same latitudes, where there is much less precipitation, and where the summer temperatures are also much higher, glaciers are absent, even on mountains 1200-1500 m. high within the Arctic circle.

Climatic zones on mountains.—The table on the next page, which is found in the volume prepared by the brothers Schlagintweit on the physical conditions of the Alps, will serve to illustrate the manner in which phenological observations emphasise the climatic zones on high mountains.

During the spring, up to the time when blossoms cease to form, the retardation in the development of vegetation amounts, in round numbers, to 10 days for every 300 m. of increase of altitude. During the fruit season, up to the beginning of winter, the retardation is 12½ days for the same distance. As the result of 10 years' observations in France, Angot found a retardation of the time of the foliation and blossoming of plants, as well as of harvesting, of 4 days for every 100 m.

CLIMATIC ZONES IN THE ALPS BETWEEN LATITUDE 46·5° AND LATITUDE 48° N.[1]

Phenomena.	Height above Sea-Level (meters).						
	500-650	650-1000	1000-1300	1300-1600	1600-2000	2000-2300	2300-2600
Melting of Snow; Awakening of Vegetation,	Mar. 17	Mar. 30	Apr. 10	Apr. 21	May 12	June 2	June 28
Cherries blossom,	May 5	May 10	May 16	May 21	June 21[2]	July 11[2]	July 29[2]
Hay Harvest,	June 15-20	June 24	June 25	June 27	July 1	Aug. 3	—
Cherries ripen,	June 25	July 18	Aug. 3	Aug. 20	—	—	—
Winter Wheat ripens,	July 18	July 31	Aug. 8	Aug. 18	Sept. 3	1690	meters.
Oats ripen,	Aug. 14	Aug. 27	Sept. 5	Sept. 16	Sept. 29		
General Snow-cover; Beginning of Winter,	Dec. 10 (?)	Nov. 30 (?)	Nov. 20	Nov. 10	Oct. 28	Oct. 15	Oct. 1

[1] Without the Western Alps. [2] Beginning of the blossoming of the rhododendron.

of increase of altitude. The arrival of swallows and the call of the cuckoo are delayed only two days in the same distance.

Schindler has made some interesting studies of the zones of cultivation in the eastern Alps, and a few of his results may be considered here. The upper limit of grain cultivation in this district coincides with the highest permanently inhabited farm-houses. Above this altitude, in the region of the mountain pastures, man appears for only two or three months of the summer, as a visitor, and leads a nomadic, herdsman's life with his cattle. This is the region of the Alpine summer pastures, and of huts which are then temporarily occupied. At still greater altitudes there stretches the region of virgin pastures, where there is no interference on the part of man with any of the natural conditions of plant life.

Two zones, an upper and a lower, may be distinguished in the district in which grain is raised. The lower reaches to the altitude at which the fields are still permanently cultivated, with a rotation of crops. Farther up, there is a moister and cooler zone, in which the ground is sometimes sown with grain, and is sometimes allowed to become grass-covered, when it is used as pasture-land.

In the Tauern, the altitude at which grain ceases to grow is 1200 m. on the north side, and 1500-1700 m. on the south side, where there is less rain and a higher temperature. The inhabited Alpine chalets extend up to 1800-2000 m. In the Brenner district, the cultivated, or grain zone, reaches up to 1160 m. on the north side, and to 1350 m. on the south. Dwellings temporarily inhabited during the summer reach 1890 m. on the north side, and 1920 m. on the south side. In the mountains of the Oetzthal, we find the following limits of altitude in the cultivation of grain (in meters):—

LIMITING ALTITUDES OF GRAIN CULTIVATION IN THE OETZTHAL.

	Grain Limit.		Summer Pastures.	
	Mean.	Maximum.	Mean.	Maximum.
North Side, Oetzthal, - - -	1420	1750	2075	2330
South Side, Schnalserthal, - -	1675	1900	2110	2310

On the sunnier and drier southern side, the zone of cultivation reaches 150 m. higher than on the northern side. In the region of summer pastures there is, however, hardly any difference in altitude on the two sides. This also agrees with the relatively slight differences of temperature between the northern and the southern side of the Alps at the height of 2000 m., to which attention has

been called by the author. At this height, this difference is very much smaller than at the level of the valley bottoms and of the intermediate altitudes.[1]

For the district of the Ortler Alps, Fritzsch has determined the zones of altitude for different exposures, and finds the following means (in meters) :—

ALTITUDE AT WHICH VARIOUS CROPS ARE GROWN ON THE ORTLER ALPS.

	Maximum.	Minimum.	Mean.
Altitude of Permanent Habitations, -	1660 S.W.	1150 N.	1380
Grain Limit, - - - - - -	1640 S.W.	1210 N.W.	1390
Mown Fields, - - - - - -	2110 S.W.	1470 N.	1770
Herdsmen's Cottages, - - - - -	2150 S.W.	1760 N.	1950
Forest Line, - - - - - - -	2160 S.W.	2100 N.	2120
Shepherds' Huts, - - - - - -	2340 S.W.	2100 N.	2190
Tree Line, - - - - - - -	2320 S.W.	2170 N.E.	2250
Orographic Snow-line, - - - - -	2750 S.	2530 N. & N.W.	2630
Climatic Snow-line, - - - - -	3090 S.	2850 N.E. & N.	2960

The means are obtained from estimates for the eight different exposures, but only the extremes are given here.[2]

On San Francisco Mountain, a volcanic peak which rises to a height of 3900 m. above sea-level in the north-central part of Arizona, Merriam has discovered a series of seven distinct zones.[3] These zones are encountered in the ascent from the hot and arid desert of the Little Colorado to the cold and humid summit of the mountain, and each of them is characterised by the possession of forms of life which are not found in the others. The limiting altitudes of these zones are as follows :—first, the arid desert, below 1800 m.; second, the Piñon belt, from 1800 to 2100 m.; third, the Pine, from 2100 to 2500 m.; fourth, Douglas fir, from 2500 to 2800 m.; fifth, Engelmann's spruce, from 2800 to 3500 m.; sixth, a narrow zone of dwarf spruce; and seventh, the bare, rocky summit, which is covered with snow during the greater part of the year.[3] These normal altitudes are averages for the northwest side of the mountain. On the southern and southwestern

[1] F. Schindler: "Culturregionen und Ackerbau in den Hohen Tauern," *Zeitschr. d. deutsch. u. oesterr. Alpenvereins*, XIX., 1888, 73-82. J. Hann: "Die Temperaturverhältnisse der oesterreichischen Alpenländer," III., 110; and *Zeitschr. d. deutsch. u. oesterr. Alpenvereins*, XVII., 1886, 62.

[2] M. Fritzsch: "Höhengrenzen in den Ortler Alpen," *Wissenschaftl. Veröffentlichungen d. Ver. für Erdkunde zu Leipzig*, 1895, II., 105-295, with chart. Also Hupfer: "Die Regionen am Aetna," *ibid.*, 293-362, with chart.

[3] C. H. Merriam: "Results of a Biological Survey of the San Francisco Mountain Region and Desert of the Little Colorado, Arizona," *U.S. Dept. of Agriculture, Div. of Ornithology and Mammalogy, North American Fauna*, No. 3, 8vo, 1890, 6-7.

slopes the zones are carried a hundred meters or more above these limits, while in the case of northern and northeastern exposures, especially in gulches and canyons, the zones may be deflected as much as 200 or 300 m. On the southwest and northeast sides of the mountain the normal average difference in altitude of the same zone is about 275 m.

A discussion of the limiting heights in different climates does not belong here. The preceding tables are given only as illustrations of the climatic conditions which determine these limits; the subject, in all its details, belongs properly within the field of botany or zoology.[1]

Influence of mountain climates on vegetation.—A few words may here be added regarding the influence of mountain climates upon vegetation. This influence results from the decreased temperature, together with the increased intensity of insolation and increased evaporation on mountains. By means of experimental cultivation at altitudes up to 2400 m., Gaston Bonnier found that the same plants, grown under similar external conditions at greater altitudes, modify their functions in such a way that both assimilation and transpiration by daylight are increased. Thus the necessary nutriment is obtained with increased intensity during the short period of growth. The amount of sugar, volatile oils, colouring matter and alkaloids increases with increased altitude; the blossoms are more brightly coloured; the leaves are thicker, and of a darker green, the twigs are shorter and cling more closely to the ground.[2] Hoffmann took a plant (*solidago virgaurea*) from the Riffelhaus (2570 m.), and planted it at Giessen (160 m.). The plant blossomed and bore fruit at the lower level (1886-1888), seven to eight weeks earlier than the native wild varieties.

The amount of salts in the air and in rain water is less at considerable altitudes than near sea-level. On the Pic du Midi (2880 m.), Müntz found only 0·34 mg. of sodium chloride in a liter of rain water; while at the foot of the mountain, the amount was 2·5-7·6 mg. The plants which grow at considerable altitudes contain correspondingly less salt. Thus, results obtained from four varieties of plants show a ratio of 0·18 for plants on mountains to 0·47 for those below. Even the milk and the blood of animals contain less salt at some distance above sea-level.[3]

[1] F. Ratzel: "Höhengrenzen und Höhengürtel," *Zeitschr. d. deutsch. u. oesterr. Alpenver.*, XX., 1889, 102-135. This is an instructive treatment of the subject with which it deals, and the reader is referred to it for more detailed information.

[2] G. Bonnier: "Cultures expérimentales dans les Hautes Altitudes," *Comptes Rendus*, CX., 1890, 363-365; "Influence des Hautes Altitudes sur les Fonctions des Végétaux," *Ibid.*, CXI., 1890, 377-380; "Des Plantes de la Region Alpine et leurs Rapports avec le Climat," *Ann. de Geogr.*, IV., 1894-95, 393-413.

[3] A. Müntz: "Sur la Répartition du Sel marin suivant les Altitudes," *Comptes Rendus*, CXII., 1891, 447-450.

CHAPTER XVIII.

MOUNTAIN AND VALLEY WINDS AND CORRELATED PHENOMENA.

Mountain and valley winds.—Mountains both give rise to certain independent, local air currents, and also modify in many ways the general movements of the atmosphere. The most important, as well as the most interesting phenomenon of the first group of these effects is the occurrence of regular day and night winds, especially in valleys. Unless the general air movement over the region is too strong, there is observable in all mountainous districts a wind blowing up the valley by day, and down the valley by night. The regularity and the velocity of these diurnal and nocturnal winds depend upon the topography and upon the temperature conditions.

"These currents of air are best developed in valleys, although they are not limited in their occurrence to these depressions; for they appear on all slopes; and the wind blowing in the valleys is only the result of oblique ascending movements by day, or of lateral downflows at night. The transition from the ascending to the descending movement is more rapid in narrow, short, canyon-like valleys, and slower in broad, open valleys, where the ascending movement is usually not fully developed until toward 10 A.M.; and where the descending night wind does not begin to blow regularly until toward 9 P.M. The hours of transition vary with the seasons. The configuration of the upper portion of a valley has a marked influence upon these winds. This influence varies with the hour of the day and with the season. Thus, the winds are sometimes more marked by day than by night; and then again they are sometimes stronger by night than by day. Occasionally, the winter, with its snowfall, is most favourable to the development of the night winds; while, on the other hand, the day winds are generally strongest

in summer." Thus Fournet, who made the first thorough study of the mountain and valley winds of the western Alps, summarises his views.[1]

Names of mountain and valley winds.—These day and night winds frequently have special names. Thus, on the Lake of Como, the wind which blows up the valley, toward the head of the lake, is called *la breva*, the names *breva di Lecco* and *breva di Como* being used with reference to the two branches of the lake. The night wind, from the opposite direction, is called *tivano*. On the Lake of Garda, during the summer, the *ora* blows as a southerly wind between 10.30 A.M. and 3 P.M., from the lower to the upper end of the lake. In the lower valley of the Etsch, as for example, at Ala, the *ora* also blows with great regularity and with high velocity up-stream during the day. The night wind on the Lake of Garda is known as *sover* (also *sopero*; in Torbole, it is called *paesano*). It blows with less regularity and with less velocity than the *ora*; but at Riva it occasionally resembles a gale. In the interval between the two opposing winds a calm prevails. On the lakes of the Austrian Salzkammergut, these winds are known by the characteristic names of *unterwind* (the day wind) and *oberwind* (the night wind). With the former, boats sail to the upper end of the lake, and with the latter they come back again at night. In the Alpine valleys, there is a well-known weather proverb to the effect that when the regular diurnal and nocturnal winds do not blow, a change to stormy weather is indicated. There is reason in this proverb, for when these winds fail, it shows that some well-defined general movement of the atmosphere prevents the development of local winds, and such conditions as these usually give rise to cloudy and rainy weather among mountains.

Mountain and valley winds in different countries.—"The winds in the mountains of the Himalayas," says Strachey, "blow up the valleys during the day from 9 A.M. to 9 P.M., and down them during the corresponding hours of the night. At the debouches of the principal streams into the plains, these night winds blow with great violence, particularly in the winter. They diminish in force as we ascend the mountains, and at great elevations, and in the plains of western Tibet the nights are almost always perfectly calm. The diurnal winds, on the other hand, in the latter country, are terrific, and in travelling there

[1] J. Fournet: "Des Brises de Jour et de Nuit autour des Montagnes," Met. du Bassin du Rhone, III., translated in *Pogg. Ann.*, *Ergänzungsband*, I., 1842, 490-511, 595-631.

we looked forward to the afternoon, when the winds are at their height, with real dread."[1]

Travellers on the great plateau of Tibet likewise find there the regular alternation of day and night winds. Of the plateau of the Karakasch, between the Kuenlun and Karakoram Mountains (latitude 34·5° – 36° N.; altitude about 5000-5500 m.), Henderson says that a strong wind blows from the west or southwest every day. This increases to the velocity of a gale in the afternoon, but calms down at night. A very strong wind blows up the great Serafshan valley from the southwest during the day. During the night it is calm, or else there is a wind from the east, which is the nocturnal wind coming down from the mountain. In regard to Java, Junghuhn reports that after 6 or 7 P.M. a wind regularly blows from the summits down the slopes of all the high mountains.[2] In many cases the night wind alone is observed, because, owing to its low temperature, it is more noticeable than the day wind. Sometimes, also, the former is stronger than the latter.[3]

At Freiburg, in Breisgau, the strong cold mountain wind which blows out of the Höllenthal through the night is often unpleasantly noticeable,[4] while the valley wind of daytime attracts no attention.[5] The latter, however, appears distinctly in the following summary of the mean frequency of the different winds at Freiburg :—

FREQUENCY OF WINDS AT FREIBURG IN BREISGAU (PER CENT.).

Hour.	N.	N.E.	E.	S.E.	S.	S.W.	W.	N.W.	Calm.
7 A.M., - -	11	4	5	**22**	10	21	5	11	11
2 P.M., - -	14	3	2	4	7	24	13	**29**	4
9 P.M., - -	5	4	9	**31**	9	21	4	9	8

[1] R. Strachey: *Proc. As. Soc. Bengal*, 1871, 16, and in S. A. Hill: *Meteorology of the Northwestern Himalayas*, 397. H. F. Blanford: *The Indian Meteorologist's Vademecum*, Calcutta, 1877, I., 167, Fig. 8. John Eliot: "A Discussion of Anemographic Observations recorded at Simla during the Period September, 1893, to August, 1896, and at Darjeeling during the Period April, 1885, to December, 1896, and an Investigation into the General Features of Air Movement in the Himalayan Area," *Indian Met. Mem.*, VI., Pt. VI., Calcutta, 1900, 445-536.

[2] See also recent account by J. H. F. Kohlbrugge: "Meteorologische Beobachtungen zu Tosari (Java)," *M.Z.*, XVI., 1899, 68-69.

[3] G. Hellmann: "Typisches Beispiel von Gebirgswinden," *M.Z.*, I., 1884, 284-285.

[4] This valley (Höllenthal) runs in a southeasterly direction from Freiburg.

[5] C. Schultheiss: "Ueber einige Eigenthümlichkeiten des Klimas von Freiburg i. B.," *Das Wetter*, XIII., 1896, 131-134, 149-153.

This table also emphasises the advantage of summarising the wind frequencies according to the different hours of observation, as noted on page 69. The wind blowing out of the Höllenthal is most frequent on spring and summer evenings (9 P.M.); the day wind from the north-west is most frequent in spring and autumn.

SEASONAL FREQUENCY OF SOUTHEAST AND NORTHWEST WINDS AT FREIBURG IN BREISGAU (PER CENT.).

	Southeast.			Northwest.		
	Morning.	Afternoon.	Evening.	Morning.	Afternoon.	Evening.
Winter, - - -	24	4	23	8	25	10
Spring, - - -	19	4	31	11	31	10
Summer, - - -	18	4	41	12	23	6
Autumn, - - -	25	3	28	11	29	9

The effect of the topography in the vicinity of Ithaca and of Utica, New York, upon the wind directions at those stations has been considered by Turner.[1] Ithaca is situated at the southern end of Cayuga Lake, the trend of the valley and of the lake being about north and south. The night, or valley wind, "usually commences from one to two hours after sunset, blowing from the south down the channels of the two principal streams flowing into Cayuga Lake. At first a light breeze, it increases in force during the night, and attains a maximum velocity probably not less than 13 km. (8 miles) per hour. The current in the main valley at the head of the lake (as observed by means of small balloons) is from 15 m. (50 ft.) to 30 m. (100 ft.) in depth before midnight, and no doubt becomes greater before morning. This volume of cold air gradually increases until sufficient to overcome the heating effect of the lake waters, reaching the northern extremity of the valley toward morning." At Utica, the valley opens both eastward and westward from the city, the highlands rising mainly towards the northeast and southeast of the city. "The prevailing winds at midday, while mainly due to the general atmospheric circulation, must be considerably strengthened by the updraught of air

[1] E. T. Turner: "The Climate of the State of New York," *Fifth Ann. Rept. Met. Bureau and Weather Service of the State of N.Y.*, Albany, 1894, 388-390.

on the heated hill slopes. At night, when the motion of the upper currents is no longer imparted to the surface air by convectional action, the downflow from the hills proceeds unchecked; but owing to the distance of the city from the highlands, the easterly wind does not become fully established there until after the evening observation, and is much more apparent in the early morning."[1]

In the valley of the river Jordan, el Ghôr, the winds are always from the north in winter, and from the south (valley winds) in summer. The east and west winds which blow on the Syrian plateau are not felt in the Jordan valley.

The wind at the mouth of the great Münster valley, in Alsace, blows out of this valley every evening after warm, calm days. It lasts throughout the night, and its cooling effect extends to considerable distances over the plains of Kolmar. A similar night wind is that of Nyons, in the department of the Drôme, in France, which since early times has been known by the name of *pontias*. This wind blows out of a deep, narrow, winding ravine nearly 15 km. long, whose mouth opens onto the plains of the Rhone, while its upper end merges into a broad valley. The *pontias* blows in summer after 9-10 P.M.; but in winter it begins as early as 6 P.M. It continues throughout the night, increasing in velocity until sunrise, when it decreases again, and after a few hours stops entirely. It is much colder and more violent in winter than in summer. The wind is also stronger when the ground is covered with snow, while it often fails altogether during the short, warm nights of summer, or when the night is rainy or cloudy. In the Rhine valley, the *Visperwind* is well known. This wind blows as a cold stream of air out of the Visp valley, which enters the Rhine valley at the village of Lorch. The long Visp valley has a great many narrow side valleys, and numerous gorges opening into it, and in these the temperature is often more than 10° lower than in the valley of the Rhine. The *Visperwind* blows only in clear weather, especially during the warmer months. It begins during the night; frequently also in the evening; and lasts until towards 9, or occasionally even 10 o'clock in the morning. In the spring, this cold wind occasionally injures the blossoming fruit trees and vineyards of the Rheingau. It frequently causes fogs in the warmer moist air over the Rhine.

During the daytime there is a wind blowing up the Visp valley; but no name has been given to this up-stream wind, nor is it known

[1] See also F. B. White: "Topographic Influence on the Winds of the Weather Maps," *Amer. Met. Journ.*, XII., 1895-96, 15-19.

among the inhabitants, although, according to Berger, it is distinctly observable.[1]

Rein has called attention to the importance of mountain and valley winds in the distribution of plants. In the case of volcanoes, the plants spread from the valleys up toward the summits, the valley wind carrying the seeds in that direction.[2]

Theory of mountain and valley winds.—The theory of these winds, which change their direction from day to night, and which in many respects are very similar to the land and sea breezes that occur along shore, has only lately been fully developed. The cool night wind is easily explained. It results from the natural slope of the surface, which carries the cold air in the valley bottoms down-stream. This wind must attain its greatest velocity where narrow, canyon-like valleys open into well-warmed broader valleys, or plains. The former valleys are colder than the latter because they have several hours less insolation. Their forest covers and their greater humidity also help to cool them. In exceptional cases, when the conditions are very favourable, the night wind may last throughout the day.

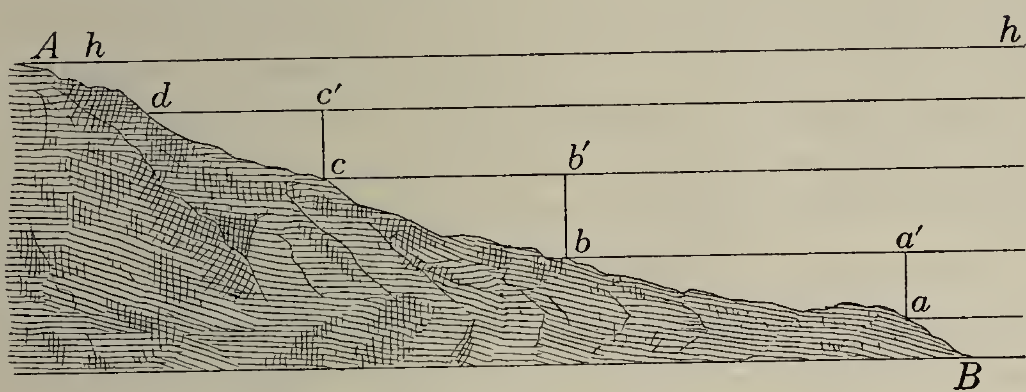

Fig. 11.—Isobaric Surfaces over a Mountain Slope.

It is not quite so clear why the air which has been warmed on the valley floor during the day does not rise vertically, instead of following the gradual slope of the valley floor, and blowing up-stream almost as a horizontal movement of the air. The accompanying figure (Fig. 11) shows the effect of mountains in drawing air up their slopes during the daytime. This process begins just as soon as the air is warmed, and affects the air over a considerable area, and at all altitudes. Hence, since the warming increases during the day, horizontal movements of the air toward the mountains are

[1] Berger: "Der Wisper- und der Bodenthal Wind," *Pet. Mitt.*, IX., 1864, 201-205. "Theorie der Berg- und Thalwinde," *Z.f.M.*, IX., 1870, 481-490.

[2] J. Rein: "Ueber Berg- und Thalwinde und ihre Beziehungen zur Vegetation vulcanischer Gebirge," *Z.f.M.*, XIV., 1879, 99-102.

brought about. Let *AB* be a mountain slope. The line *hh* and the other lines parallel to it are horizontal. When the temperature is about the average and there is no general movement of the atmosphere over the district, the pressure is the same at all points along every one of these lines. Hence no reason for any movement of the air exists. When the sun rises, the whole mass of air in the valley and over the mountain slope is warmed. The effect of the increasing temperature is an increasing expansion of the air. This results in disturbing the equilibrium, and hence the air must flow toward the mountain side. The column of air, *aa′*, is expanded by reason of the increased temperature; and the pressure at the point *a′* therefore rises, because some of the air which before was below *a′* has been raised above that point, and hence increased the pressure at *a′*. The pressure at *b*, on the side of the mountain, however, remains constant, because it is on the same horizontal line. The same is true of the point *c*, as related to *b′*; of *d*, as related to *c′*, etc. In other words, along every horizontal line the pressure increases with increasing distance from the mountain slope, while it remains constant on the slope itself.[1]

Thus the isobaric surfaces are now no longer level, but slope toward the mountain, and the air at all altitudes descends toward the latter. If, at the same time, the mountain slope is itself warmed by the sun, the air at that slope is warmer than the free air at the same altitude (*e.g.*, the air at *b* is warmer than at *a′*), and it therefore has a tendency to rise. Thus there are two forces which determine the movement of the air along the mountain sides—a force which acts horizontally, and one which acts vertically. These two forces together cause the air to ascend along the mountain slopes by day; while the air above the valley or the lowland actually flows toward the mountain horizontally. In drawing the surrounding air towards them during the daytime, mountains may therefore be said to act like local, stationary barometric depressions.[2]

Down-cast diurnal winds from glaciers and snow-fields.—When the mountain sides are colder than the surrounding air, cold winds may blow down onto the lowlands, even during the daytime. Such winds are regularly observed at the foot of glaciers on warm days.

[1] The expansion of each column of air by the increased temperature is proportional to its depth, and therefore decreases toward the mountain summit.

[2] In regard to the theory of mountain and valley winds compare also E. Chaix: "Théorie des Brises de Montagne," *Le Globe* (Geneva), *Mémoires*, XXXIII., 1894, 105-133.

Cold down-cast winds of this kind are described by Moritz Wagner [1] as occurring on the plateau of Quito, which is surrounded by snow-covered volcanoes. "As the snow-line is approached," says Wagner, "another phenomenon is noted. This attains such indescribable violence during certain months, and in certain districts, that mountain-climbing becomes difficult, and often quite impossible. I refer to the severe storms of the *Nevados*, whose icy-cold winds blow with great violence down to the warmed valley bottoms, chiefly in the months of August, September, February, and March. These winds are often so strong that they become dangerous. The greater the accumulations of snow on the different gigantic volcanoes, and the more extended and the more barren the plateaus and elevated valleys upon which the sun shines from an almost cloudless sky during the four months named, the more regularly and the oftener do these cold storm winds blow. During my stay in Guaranda in September, 1858, a whole week passed by during which no caravan dared to cross the pass leading around Chimborazo to Chuquipoyo. Down from the slopes of this mountain, in a southeasterly direction, the snowy wind raged with indescribable fury until three o'clock in the afternoon.

"These storms usually begin when the regular thunderstorm season is over and when the sun is near the tropics, and they noticeably increase in intensity when a good many clear days follow one another without interruption. The wind begins, as a rule, about 7 A.M.; increases in velocity with the increasing altitude of the sun, and ceases to blow after sunset."

Reid has described the occurrence of local winds at the foot of the Muir Glacier, on the coast of Alaska.[2] The prevailing wind on the Alaskan coast is from the southwest, but the glacier, by cooling the air in contact with it, gives rise to a cold wind which blows down the slope. At the camp occupied by Reid, a northeast wind blew continuously, except when a strong southerly gale overcame it. On the western tributary of the glacier there was a west wind. Everywhere the local wind blew down the slope of the glacier.

Similar winds have been observed by Scott Elliot on Ruwenzori, in equatorial Africa. "In those valleys which lead directly to the base of the snow peaks . . . an extremely violent wind (almost a hurricane) blows down from the mountain from about 6 to 7 (in the evening), then dying suddenly away to nothing. This is simply the cold air from the snow rushing down to the heated lower slopes. This wind does not occur on evenings when there has been rain on the lower parts, which is what one would expect. Sometimes at the same moment there are, in an upper current of the atmosphere, clouds moving towards the upper peaks."[3]

Valley wind in the upper Engadine.—Another apparent exception to the theory that the valley wind blows up-stream is noted in the upper Engadine. Here the wind blows down the valley of the Inn from the Maloja Pass during the warmer months, while the side valleys have normal valley winds blowing up-stream. This apparent anomaly

[1] M. Wagner: *Naturwissenschaftliche Reisen in Südamerika*, 555.

[2] H. F. Reid: "Studies of Muir Glacier, Alaska," *Nat. Geogr. Mag.*, IV., 1892, 19-84.

[3] G. F. Scott Elliot: "Expedition to Ruwenzori and Tanganyika," *Geogr. Journ.*, VI., 1895, 301-317.

has been explained by Billwiller, and is actually found to confirm, in the most striking manner, the theory of valley winds which has been discussed above. The upper Engadine valley is not enclosed at its upper end, the Maloja Pass being but little higher than the valley floor itself. On the further side of the Pass, however, is the head of the deep valley of the Maira, the upper Bergell, which is well warmed. The air in this valley, which is really partly in Italy, is thus well warmed, and it rises above the level of the Maloja Pass, descending the valley of the Inn. Simultaneous observations of the air pressure at Maloja, in the upper portion, and at Bevers, in the lower portion of the upper Engadine valley, have actually shown a gradient from Maloja toward Bevers by day and a reversed condition at night.[1]

The following figure (Fig. 12) shows the rise in the isobaric surfaces

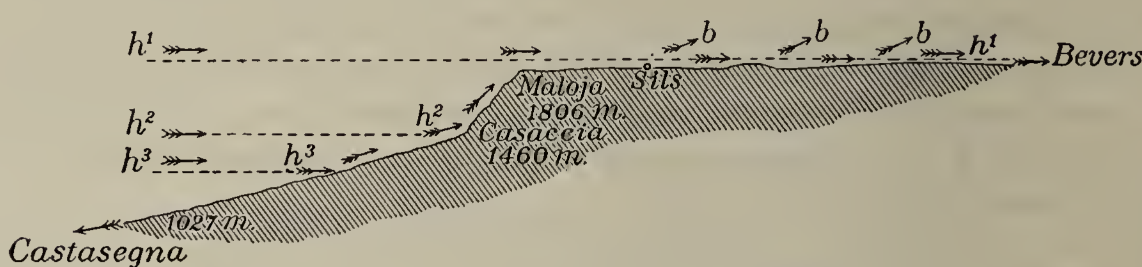

Fig. 12.—Valley Wind in the Upper Engadine.

over the Bergell, and the origin of the valley wind which flows down-stream in the upper Engadine valley by day. The wind arrows, *b*, *b*, *b*, show the normal ascending diurnal winds in the adjoining tributary valleys.[2]

In the discussion entitled "Weitere Untersuchungen ueber die tägliche Oscillation des Barometers," *D. W. A.*, LIX., 1892, 333-337, the author of the present book has tried to answer the objections to his theory which have been urged by Sprung. Moreover, the valley wind of the upper Engadine is itself the best answer to these objections. See R. Billwiller: "Der Thalwind des Oberengadin." *Ann. d. schweiz. Met. Centr.-Anstalt.*, 1893. Summarised in *M.Z.*, XIII., 1896, 129-138.

The origin of the night wind may best be studied by referring again to Figure 11, page 333. When the sun has set, the air, especially that nearest the ground, cools by radiation. The atmospheric strata, which were previously expanded, now contract. The air-columns, *aa'*, *bb'*, etc., become shorter, *i.e.*, the pressure at *a'* falls as compared with that at *b*, and the same thing happens at *b'* as compared with *c*, etc. The slope of the isobaric surfaces gradually changes again, first to the normal, horizontal condition, which results in a calm, and then to a condition in

[1] R. Billwiller, *loc. cit.* J. Hann: *Lehrbuch der Meteorologie*, 1901, 438, gives a sketch of the topography of this district (Fig. 43).

[2] R. Billwiller: "Der Thalwind des Oberengadin," *Z.f.M.*, XV., 1880, 297-302.

which they slope from the mountains towards the lowlands. Thus the air finds a slope from the mountain side out into the open. Now since the cooling of the earth's surface is greater than that of the free air at a distance from the mountain, the cooler, heavier air on the mountain slopes and valley bottoms flows down into the valley, and then down stream. Thus the cool, descending mountain winds of night-time are produced.

On the Atter See, in the Salzkammergut, brisk southerly winds blow over the lake and along its shores on fine summer mornings. These are the descending mountain winds of night-time, and they produce well-marked waves. But at a distance of only a kilometer from the shore it is nearly calm, and while it is then often quite cool on and close to the lake, it is very warm at this distance back from the lake. The houses situated directly on the shore often have cool windy nights and mornings, while these conditions are not found a short distance inland. This must frequently occur in the case of lakes enclosed by mountains. The inequalities of the mountain slopes, and trees, interfere with the movement of the air along the shores. The day wind comes from down the valley about noon and in the afternoon. The greater velocity of the wind over the smooth surface of the lake is characteristic.

Effect of local accumulations of snow upon mountain and valley winds.—The local cooling effect upon the air which is exerted by masses of snow lying in shady ravines, may cause the warmer air of the slopes which are not snow-covered to descend after sunset. The flowing off of the cold air below likewise draws down the warmer air from aloft by a kind of suction effect. On the other hand, mountain slopes which are free from snow may, by their own warming, cause the cooler air of the snow-covered valley bottom to ascend during the daytime. Pittier has made some interesting observations on this point.[1]

Diurnal variation of humidity, cloudiness, and precipitation on mountains.—The periodic alternation of ascending and descending winds in mountain regions is of the greatest importance in mountain meteorology, especially in connection with the diurnal variation of humidity, cloudiness, and precipitation. The currents of air which rise along the mountain slopes by day carry up with them the water vapour of the lower strata of the atmosphere. Hence the relative humidity aloft increases in the afternoon, while it decreases in the valleys below. Water vapour from the surrounding lowlands is thus collected during the afternoon hours above all considerable mountain masses from which numerous valleys radiate. The cooling which results from this ascending air movement condenses this water vapour into clouds. These form at some distance above the mountain summits in

[1] Pittier: "Note sur les Vents de Montagne," *Bull. Soc. Vaud.*, XVI., 604.

dry weather, but when the air is damper they envelop the mountain tops, and not infrequently develop into showers or thunderstorms.

There is a tendency to afternoon rains among mountains, and during the warmer months, to thunderstorms. Such rains occur even when the general weather conditions do not indicate any precipitation, and when the surrounding lowlands are having the most beautiful weather. These thunderstorms remain within the mountain district in which they originate; they break up toward evening, and are followed by a clear night.

The descent of the air at night, on the other hand, carries water vapour down from aloft to lower levels; the clouds dissolve, and the air on the mountains becomes drier. This was observed by de Saussure many years ago on the Col du Géant, and it surprised him greatly. For this reason the best view from mountain tops may be obtained in the early morning, because at that time the damper air is below, and there is no wind. In the afternoon, on the other hand, the warm ascending currents make the atmosphere hazy. It may even be cloudy at greater altitudes, and a bluish haze may cover the distant landscape.

There is, however, another cause for the decreased transparency of the air on mountains in the afternoon. During the afternoon the warming of the earth's surface gives rise to ascending and descending currents of air, so that the atmosphere immediately surrounding the earth is traversed by vast numbers of warmer and colder air currents, moving past one another up and down. Thus each stratum becomes an anisotropic turbid medium. This does not happen in the early morning; for then all the atmospheric strata are still isotropic, at least in a horizontal direction, and that is the important point.

The diurnal march in the growth of clouds on the Faulhorn has been admirably described by Bravais, who notes the ascending movement of the clouds during the day, their disappearance in the evening, and their reappearance as fogs in the valleys at night. They then lie in horizontal layers over the valleys and remain, "as if deprived of motion and of life." This quiet at night is in striking contrast to the ascensional movement of the accompanying clouds over the mountains by day. Then they are in active motion throughout their mass, as is shown by the development of vortex rings within them, and by their repeated formation and dissolution. Bravais gives an excellent description of the way in which the sun produces movements in the lower strata of the atmosphere.[1] A similar description of the diurnal changes in the clouds seen from Mont Blanc has been given by Janssen.[2]

[1] A. Bravais: *Mémoires sur les Courants ascendants de l'Atmosphère*, Lyons, 1842.

[2] Translated in *Am. Met. Journ.*, XII., 1895-96, 354-355.

The daily migration of the water vapour upward along the slopes to the mountain tops where condensation takes place during the afternoon, and back again to the lower levels at night, has been graphically described by Junghuhn in the case of Java. Junghuhn's account is typical for all mountain regions in warm countries. The phenomena described are not, however, so well developed, nor do they occur so regularly throughout most of the year in all parts of the world, as is the case near the equator.

"We are in the interior of Java, on the southern slope of the northern Bandong range, at about 1200 m. above sea-level, with an outlook over the whole Bandong plateau (700 m.), and its enclosing mountains of 1600 to 2600 m. in altitude. The evening is clear. The moon shines brightly on the plateau; not a breath of air is stirring; and not a cloud, nor a trace of fog is to be seen. The night passes on peacefully. On the following morning we again gaze downwards, and seem to be looking at the surface of a great lake in the depths below. The whole plateau is covered with a sheet of fog, whose even upper surface is greyish-white at first, but becomes a dazzling white as soon as the rays of the sun fall upon it.[1] Half an hour later, the surface, which at first was perfectly level, begins to rise in small waves; to surge to and fro, and finally to gather in masses as cumulus clouds. The sheet of fog breaks up more and more into single masses of cloud, which rise higher and dissolve, and between 8.30 and 9 in the morning both fog and clouds have usually wholly disappeared. Then clouds appear over the slopes of the mountains which surround the plateau, which up to this time have been entirely cloudless, like the rest of the sky. Against the dark background formed by the virgin forests of these great slopes, detached cloudlets may first be discovered. These appear so suddenly that they at first seem to be clouds of steam, rising from the crater of a volcano. Soon, however, their numbers increase so rapidly, and they all form close to one another in a horizontal line at so constant a height above the surrounding mountains, that the mistake is detected. These clouds visibly increase in size. They grow, unite, and finally, by about 10 or 10.30 in the morning, they form a strato-cumulus cloud, whose sharply-defined base looks as if it were cut off along the

[1] The clearer the night, the more extended and the deeper is this sea of fog. Under these conditions, the highest trees (16—23 m.) alone project above it, appearing like dark rocks or little islands. The fog does not begin to form until 2 or 3 o'clock over the lowest portion of the plateau, but by 4 o'clock the whole plateau is covered.

mountain sides, while its upper surface is in constant motion, tossing about with a wave-like movement, and rising in towering masses.[1]

"Over the central portion of the plateau itself the sky remains clear with the exception of a few white cumulus clouds, but the heated air is by no means clear, being filled with a whitish, cloudy haze. On the mountains which are nearest the damper, northern coast, on the other hand, the strato-cumulus clouds form a single, continuous sheet as early as 2 P.M. This cloud cover keeps growing darker and heavier near the mountains, until at about 3 or 4 P.M., and on the mountain sides themselves by 2 P.M., the rolling thunder tells of the electrical discharges which are taking place among the clouds.[2] Then places which are situated near these mountains, like Buitenzorg, for example, have very heavy thunderstorms almost daily, while over the central plateau of Bandong not a drop of rain falls, and on the surrounding mountains there is only an occasional clap of thunder over some forested peak.

"Hardly has the sun sunk below the horizon, than we notice to our surprise that all the clouds which were floating in the air have disappeared. Indeed, if we look over towards the dense clouds which wholly concealed the summits and upper portions of all the mountains after noon, we note with increasing wonder, that these clouds are visibly dissolving. They become smaller and smaller before the eyes of the observer, and even before the last ray of daylight has faded away from the sky, there is not a trace of them left.[3]

"There is not the slightest breath of air; no fog is visible on the plateau, and the disappearance of all the great masses of cloud from

[1] This cloud belt has an average altitude of about 1500 to 2400 m. in Java. During the dry season, its lower limit is higher (1800-1900 m.), and its thickness is only a few hundred meters.

[2] In regard to the altitude at which these thunderstorms develop, Junghuhn says elsewhere: "At altitudes of 3000 m. and more, thunderstorms very seldom occur, and then only during the rainy monsoon. Standing on the side of these lofty peaks in the afternoon you will generally be in the brightest sunshine, while 1000 to 1600 m. lower down, on the sides of the mountain, the lightning is flashing through the clouds and the thunder is pealing." In higher latitudes, as, for example, in the Alps, thunderstorms form at greater heights, and observations of this kind can but seldom be made.

[3] The local mountain thunderstorms of the Alps, after they have exhausted their energy, and their rain has stopped, leave behind them a high cirro-stratus cloud cover, which gradually becomes thinner, and through which the stars shine again at night. The sky is perfectly clear in the morning. The thunderstorms of the rainy season at Para exhibit a precisely similar succession of phenomena, according to Bates.

the air and from the mountain sides within so short a time, seems almost magical. On the driest days the clouds disappear soon after sunset, and on other days before midnight. The night is perfectly calm and clear. Not a trace of fog is to be seen. When, however, the moon is shining on the plateau, a sheet of fog may be seen to form about 3 A.M. on the surface of the plateau, the fog being first visible on the lowest portions of this surface. When this fog has once begun to form it spreads with extraordinary rapidity, and in less than an hour it covers the whole plateau, which is 50 km. wide. When the sun rises, it shines upon a sea of fog, just as it did on the preceding morning at the same hour. This fog later again rises into the air like steam and the changes of the preceding day are repeated."[1]

The lofty mountains of the equatorial zone and of the tropics in general, especially those which are snow-capped, are usually invisible, because they are almost always covered with clouds during the daytime in consequence of the regular diurnal ascending currents up their sides. Wagner and Whymper give graphic descriptions of this phenomenon in connection with the snow-covered mountains of Ecuador. According to the former, there are hardly 60 to 70 days in the year on which the mountain climber on the high Andean summits is not likely to encounter one of the regular afternoon thunderstorms, which are always accompanied by hail and snow. Of Illiniza, Whymper says that he could not once in 78 days see the whole mountain. Even during May and June, in the dry season, his guide, Carrel, saw the mountain but twice in five weeks.[2] During Whymper's encampment, from the end of December to the middle of January, at an altitude of about 5000 m., the snowy peaks of Chimborazo could clearly be seen during the morning hours up to 8 A.M. The clouds were then below the summit, reaching up to about 4000 m. above sea-level. After 8 A.M., the clouds began to move upward on the eastern side of the mountain, and they concealed the summit after 10 A.M. Thunderstorms occurred every day regularly, and many were extraordinarily severe. These storms seldom began before noon. Snow and hail fell every day.

Precisely similar observations have been made on the snow-covered mountains of eastern equatorial Africa.

[1] Junghuhn: *Java*, I., 288-291. Further on (p. 354), there is an interesting description of a similar diurnal variation of cloudiness and of rainfall on Sumbing, with illustrations of the cloud forms.

[2] E. Whymper: *Travels amongst the Great Andes of Ecuador*, 1896, 130.

The daily formation of a cloud cap upon Kilimanjaro attracted the attention of Rebmann. Von Höhnel notes that Kibo and Kimawensi always become covered with clouds between 8 and 9 A.M. The clouds always begin to form over the summit of the latter mountain, although it is the lower of the two and is not covered with snow. After sunset, Mt. Kibo first becomes cloudless, and then Mt. Kimawensi, for a few minutes only; but sometimes the clouds do not leave the latter mountain at all. Scott Elliot says that the most remarkable phenomenon on Ruwenzori is the white cloud which always covers the upper portion of the mountain. This cloud is at an altitude of about 2000 m. in the morning, while it extends a few hundred meters nearer sea-level in the valleys. The belt of moist virgin forest follows the mean altitude of this cloud. After 10 A.M., this cloud slowly begins to rise, and occasionally wholly disappears by 5.30 P.M. This is the only time during the day when the snow-covered peaks are visible.

The snow-covered ranges of the Himalayas can be seen from the Indian side only on winter mornings; for they are permanently enveloped in clouds during the summer rainy season.

Sir Richard Temple gives a very characteristic description of the daily sequence of weather phenomena on the plateau of the lake region in Sikkim, on the borders of Tibet, at an altitude of about 3600-4800 m. above sea-level (Chola Pass; Bhewsa Lake). "Early in the morning at sunrise, the weather is quite superb; the sky is unclouded azure, and the mountains are unbroken dazzling white. This lasts for some three hours, that is to say, till about 10 o'clock in the day, and that is what we used to call the bloom of the morning. Then, and then only can you take your sketches. The air is extremely cold—biting cold. After 10 o'clock up come the clouds. You cannot tell how they form themselves. A little bit of vapour, no bigger than a man's hand, expands; fresh men's hands arise and clouds accumulate, till at last the whole atmosphere is clouded over. This lasts till about middle day. Then the clouds seem to turn into snow, and a certain amount of snow falls in the afternoon, which makes you very miserable in the evening, and you sit down to dinner with snow all around you, and your little tent also encrusted with snow. But towards midnight the clouds pass away, and stars come out, and it is a magnificent night. Then you have the sunrise as already described."[1] It is hardly ever possible to get a view of the mountains at sunset. Temple saw Kanchanjanga only once in the light of the setting sun.

All mountainous districts in warmer latitudes show this diurnal variation of cloud formation and of precipitation, although in higher latitudes these phenomena are noted only in summer, and during spells of calm, hot, damp weather. The afternoon thunderstorms, which occur day after day in the Alps,[2] the Rocky Mountains,[3] on the plateau

[1] Sir R. Temple: *Oriental Experience*, London, 1883, 81-82.
[2] J. Hann, *Z.f.M.*, VIII., 1873, 102-106.
[3] C. C. Parry, *Am. Journ. Sci.*, 2nd Ser., XXXIII., 1862, 231-237.

of Costa Rica,[1] and in the Blue Mountains of Jamaica,[2] all agree in their daily periodicity. This depends upon the transportation of water vapour aloft by the ascending diurnal winds. In all these districts, the daily thunderstorms are followed by clear nights and mornings.[3]

[1] Moritz Wagner, *Ausland*, 1854, and *Naturwissenschaftliche Reisen im tropischen Südamerika.*

[2] Arago's works, Vol. IV., where many older observations may also be found.

[3] The description given by Desirée Charnay, of the daily sequence of weather changes at Amecameca, at the base of Popocatepetl, 200 m. above the valley of Mexico, is very characteristic. "Le matin, tout est calme, paix, silence, beauté. Le soir, tout est bruit, colère, tourmente, lutte des éléments entre eux" (*Le Tour du Monde*, 1881, II., 288).

CHAPTER XIX.

THE FOEHN, SIROCCO, BORA AND MISTRAL.

The foehn.—Mountains locally modify, to a considerable extent, the air movement which results from the general distribution of pressure. The most important and the most instructive of these influences is undoubtedly that which was first carefully studied in northern Switzerland, where, because of the presence of the Alps, the southerly winds may develop certain special characteristics which give them the name of *foehn* winds.[1]

[1] References on the *foehn* in general:—H. W. Dove: *Ueber Eiszeit, Föhn und Scirocco*, Berlin, 1867; *Der Schweizer Föhn* (Supplement), Berlin, 1868. H. Wild: *Ueber Föhn und Eiszeit*, Bern, 1868; *Der Schweizer Föhn* (an Answer to Dove), Bern, 1868. L. Dufour: "Récherches sur le Foehn du 25 Septembre, 1866, en Suisse," *Bull. Soc. Vaud. des Sci. nat.* (Lausanne), IX., 1868. A. Hirsch: "Les Récherches récentes sur le Foehn," *Soc. de Neufchâtel*, 1868. J. Hann: "Zur Frage über den Ursprung des Foehn," *Z.f.M.*, I., 1866, 257-263; "Der Föhn in den oesterreichischen Alpen," *Ibid.*, II., 1867, 433-445; "Der Scirocco der Südalpen, *Ibid.*, III., 1868, 561-574, translated by L. Dufour, with some comments, in *Archiv. des Sci. phys. et nat.* (Geneva), XXXIV., 1869, 231-242; "Ueber den Föhn in Bludenz," *S. W.A.*, LXXXV., 2, 1882, 416-440; "Einige Bemerkungen zur Entwickelungs-Geschichte der Ansichten über den Ursprung des Föhn," *M.Z.*, II., 1885, 393-399. H. Wettstein: "Ueber den Föhn," *Verhand. schweiz. naturforsch. Gesell.*, Schaffhausen, 1873, 169. R. Billwiller: "Ueber verschiedene Entstehungsarten und Erscheinungsformen des Föhns," *M.Z.*, XVI., 1899, 204-215; "Bildung barometrischer Theilminima durch Föhne," *Ibid.*, XVIII., 1901, 1-4. G. Berndt: "Der Alpenföhn in seinem Einfluss auf Natur- und Menschenleben," *Pet. Mitt.*, *Ergänzungsheft* 83, Gotha, 1886; *Der Föhn*, Göttingen, 1886 (a summary of value to anyone who has not access to the original papers because of the quotations which it contains). Herzog: "Der Föhn: Auftreten, Erklärung, Einfluss auf Klima und Organismus," *Jahresber. St. Gall. naturw. Gesell.*, 1889-90. E. Bosshard: "Ueber Herkunft und Entstehung der Föhnstürme," *Naturforschende Gesellschaft Graubündens*,

Characteristics of the foehn in Switzerland.—The foehn is a warm dry wind, which blows down from the crest of the Alps with great violence from a southeasterly, southerly, or, less frequently, from a southwesterly direction. The exact compass-point from which the wind blows clearly depends chiefly upon the trend of the valleys. Even in places where the foehn always appears as a southeast wind, it is noted that the clouds which are visible at the same time at greater altitudes drift from the southwest. The main valleys which trend S.E.-N.W., or S.-N., on the northern side of the central range of the Alps are the most exposed to the foehn. The valleys whose trend is east and west are seldom, or never, visited by the foehn. Thus the upper Valais, and the valley of the Aar between Brienz and Thun, seldom have foehn winds; while above Meiringen and at Guttannen, where the trend of the valley is north and south, the foehn frequently blows. The district of most frequent occurrence is between Geneva and Salzburg, immediately adjoining the main chain of the Alps to the south; and the velocity of the foehn, as well as the rise of temperature and the dryness which accompany it, are greatest in the valleys themselves. The foehn attains its greatest development in the valley of the Vorarlberg Ill (at Bludenz, for example); in the valleys of the Rhine as far as the Lake of Constance; of the Linth, to near Zürich; of the Reuss and of the Engelberg Aa, to near Muri; and of the lower Rhone, to the Lake of Geneva. In the upper portion of the valleys of the Rhine, the Linth and the Reuss, as well as in the lower Rhone valley, the foehn sometimes attains the velocity of a gale. The violence decreases with increasing distance from the main range of the Alps, and throughout the greater part of the Swiss highland, in the Jura and beyond the northern frontier of Switzerland, the foehn is noticeable only because of the slight rise in temperature and the decrease in the humidity which it causes.

The phenomena associated with the occurrence of the foehn in Switzerland are thus described by Tschudi: "A light veil of cloud over the mountain summits appears on the southern horizon; the sun sets pale and lustreless in the red sky.

1894. J. M. Pernter: "Ueber die Häufigkeit, die Dauer und die meteorologischen Eigenschaften des Föhn in Innsbruck," *S. W.A.*, CIV., 2a, 1895, 427-461; "Die allgemeine Luftdruckvertheilung und die Gradienten bei Föhn," *Ibid.*, CV., 2a, 1896, 117-137. Mühry: "Ueber den Föhnwind," *Z.f.M.*, II., 1867, 385-397; "Zur Kartenskizze eines Föhnwindes," *Ibid.*, III., 1868, 363-364. H. Wild: "Ueber den Foehn und Vorschlag zur Beschränkung seines Begriffs," *Denkschr. schweiz. naturf. Ges.*, XXXVIII., 2, Zürich, 1901 (*M.Z.*, XVIII., 1901, 476-479).

The clouds glow for a long time with the most vivid purple colours. The night is oppressive, and no dew is formed. Cold breezes are noted from time to time, occurring in streaks. The moon is surrounded by a dull reddish corona. The air is exceedingly clear and transparent, and the mountains seem to be much nearer than usual. There is a bluish-violet tinge to the distant landscape.[1]

"In the distance is heard the rustling of the forests on the mountains. The roar of the mountain torrents, which are filled with an unusual amount of water from the melting snows, is heard afar through the peaceful night. A restless activity seems to be developing everywhere, and to be coming nearer and nearer. A few brief gusts announce the arrival of the foehn. These gusts are cold and raw at first, especially in winter, when the wind has crossed vast fields of snow. Then there is a sudden calm, and all at once the hot blast of the foehn bursts into the valley with tremendous violence, often attaining the velocity of a gale which lasts two or three days with more or less intensity, bringing confusion everywhere; snapping off trees; loosening masses of rock; filling up the mountain torrents; unroofing houses and barns—a terror in the land. Those portions of the valleys which are nearest to the southern mountain barrier usually have the most violent foehns."[2]

Relations of the foehn to customs, habitability and crops.—Human beings as well as animals suffer under the influence of this wind, which relaxes the nerves, and has a depressing effect upon the mind. Fires in fire-places or stoves are carefully extinguished. In many Swiss valleys fire patrols go quickly from house to house in order to be sure that the fires have been extinguished, for great conflagrations may easily occur at such times, owing to the drying of the wood by the wind. On the other hand, the foehn is also a welcome visitor, especially in spring, for it causes the snow and ice to melt rapidly, and almost in an instant changes the whole aspect of the country. In the Grindelwald valley the foehn often melts a snow-cover more than 65 cm. thick in 12 hours. The foehn is the real harbinger of spring. It accomplishes as much in 12 hours as the sun does in two weeks; for even the old snow, which has become very compact, and upon which the sun has long been shining without effect, cannot resist the foehn. Indeed, in many shady valleys high up among the mountains, this wind actually brings on spring, just as, in the autumn, it ripens the grapes in many

[1] This rich deep blue, or violet colour of the mountains and the clearness of their outline before the arrival of the foehn are noted in New Zealand, where the foehn comes from the northwest, as well as by observers in the Alps.

[2] The clouds over the mountains lying south of the valleys often have the appearance of a massive wall, and this is known by the name of the foehn-wall. The foehn-wall as it appears from the Gschnitz valley above the Stubai ranges has been described, illustrated and explained by F. Von Kerner ("Die Föhnmauer: eine meteorologische Erscheinung der Zentralalpen," *Zeitschr. d. deutsch. u. oesterr. Alpenver.*, 1892).

parts of the lowlands. In Graubünden, in particular, an abundant grape-harvest is expected at the end of August and in September, if the foehn comes at the right time and lasts long enough. Here the ripening of the grapes actually depends upon the foehn. Similarly, the cultivation of corn in Vorarlberg, and in the northern Tyrol, depends upon the frequent occurrence of the foehn, which is called the *sirocco* in Innsbruck. The small district in the valley of the Inn, which is exposed to the "warm wind" coming down from the Brenner Pass, and at whose centre lies the city of Innsbruck, according to Kerner actually represents a southern island, from a botanical standpoint. To give only one example, the flora of Innsbruck includes the hop-hornbeam, which occurs nowhere else in the northern portion of the central chain of the Alps.

Temperature and humidity of the foehn.—Stations at which the foehn frequently occurs, and where it is well developed (as *e.g.*, Bludenz and Altdorf), have a relatively mild autumn and winter temperature. The following examples may serve to illustrate the high temperature and the dryness of the foehn.

TEMPERATURE AND RELATIVE HUMIDITY DURING FOEHN WINDS AT BLUDENZ.

Date.	Temperature.			Relative Humidity (%).			Wind.
	6 A.M.	2 P.M.	10 P.M.	6 A.M.	2 P.M.	10 P.M.	
	°	°	°	°	°	°	
Feb. 16, 1867,	12·5	17·0	14·0	26	21	26	SE. 5
Jan. 31, 1869,	13·8	16·0	13·3	6	11	24	SE. 5
Feb. 1, 1869,	14·0	19·3	—	20	14	—	SE. 5
Dec. 10, 1856,	13·5	18·0	14·0	27	13	30	S. 7
Nov. 24, 1870.	—	15·0	16·5	—	12	12	SE. 4-5
Nov. 25, 1870,	17·3	22·0	—	13	10	—	SE. 5
Mar. 6, 1871,	10·7	17·2	12·5	20	9	14	SE. 5-6

TEMPERATURE AND RELATIVE HUMIDITY DURING THE FOEHN OF JANUARY 1, 4, 7-9, 1877, IN SWITZERLAND.

	Temperature.			Relative Humidity (%).			Mean Wind Direction.
	7 A.M.	1 P.M.	9 P.M.	7 A.M.	1 P.M.	9 P.M.	
	°	°	°				
Altdorf, - -	13·8	15·8	13·0	31	29	42	S.
Altstätten, -	15·1	16·0	14·0	25	29	35	SSW.

The temperature in mid-winter is thus seen to rise as high as summer, and the relative humidity is lowered to a remarkable extent. The departure of the temperature from the normal amounted to $+15 \cdot 7^\circ$ in the foehn of January 31-February 1, 1869; and the departure of the relative humidity was -58 per cent. During the foehn of January, 1877, the departure of the mean daily temperatures at Altstätten, in the Rhine valley, was $+17 \cdot 1^\circ$ on January 1, and $+17 \cdot 2^\circ$ on January 8; and it was almost as large at Altdorf. In short, as soon as it blows with any considerable force, the foehn is an extraordinarily warm and dry wind.

During the foehn of January 27, 1890, the relative humidity at Partenkirchen was reduced to 6 per cent. between 2 and 6 P.M. (minimum 4 per cent.), as determined by a psychrometer and a hair hygrometer. The temperature was $12^\circ - 13^\circ$. The snow disappeared with a rapidity which had never been known before. The skin broke open; finger nails split, and the feeling of dryness was extremely unpleasant.[1]

At Bludenz, even taking the mean of all cases during ten winters, the winds from between east and south, which blow down from the crest of the Rhætikon, and from the Silvretta range (mean relative altitude, at least 2000 m.) raise the temperature $8 \cdot 2^\circ$ above the normal; and lower the humidity 31 per cent. below the mean. The mean maximum temperatures of the months from November to February at Bludenz are higher than the corresponding extremes of stations at the southern base of the Alps, such as Milan, Riva, and Bozen; and this difference is due to the foehn winds at the former station.

Seasonal occurrence of foehn winds.—The average frequency of days on which foehn winds occur in northern Switzerland has been found by Wettstein to be as follows (7 year means):—

NUMBER OF DAYS WITH FOEHN WINDS IN NORTHERN SWITZERLAND.

Winter.	Spring.	Summer.	Autumn.	Year.
9·1	17·3	4·9	9·6	40·9

Ten years' observations gave the following results at Bludenz (Hann) and Innsbruck (Pernter):—

[1] M. F. Ward: "Remarkably Dry Air," *Symons' Met. Mag.*, XXV., 1890, 7 (*M.Z.*, VII., 1890, 240).

NUMBER OF DAYS WITH FOEHN WINDS AT BLUDENZ AND INNSBRUCK.

	Winter.	Spring.	Summer.	Autumn.	Year.
Bludenz, - - -	10·6	8·2	3·1	10·0	31·9
Innsbruck, - -	9·5	17·0	5·0	11·1	42·6

Foehn winds occur least frequently in summer, and are not so well developed then, while they are most frequent in winter and spring. On the average, each year has between 30 and 40 days with foehn winds. It is, therefore, easy to see that the foehn must also have a marked effect upon the mean temperature of the places at which it frequently occurs. Pernter has determined the effect which the foehn winds have upon the mean temperature at Innsbruck. He finds that they raise the mean temperature 0·8° in winter and spring; about 0·2° in summer; and 0·7° in autumn. The mean annual temperature at Innsbruck is raised 0·6°, and this corresponds to a change of latitude 1° further south.[1]

Theory of the foehn.—In seeking an explanation of the foehn, it was at first naturally enough supposed that this wind came to Switzerland from northern Africa. The wind is hot and dry; it comes from the south; and south of Switzerland lies the Sahara, which is the very embodiment of heat and aridity. It might, to be sure, have been remembered that the Sahara is very hot in summer only, while the foehn causes the greatest rise of temperature in winter, and is least well-developed in summer.

The theory which holds that the foehn comes from the Sahara is emphatically contradicted by the character of the weather which is noted on the northern and southern sides of the Alps while the foehn is blowing. While the hot, dry, southerly winds are blowing in northern Switzerland there is a calm on the southern side of the Alps; the temperature in the valleys is increased little, if at all; the relative humidity is high. In most cases also, a few hours after the foehn has begun to blow in the northern Alpine valleys, rain and snow begin to fall on the southern slopes and on the summits of the Alps. This precipitation is often extraordinarily heavy. During the terrific rains and

[1] J. M. Pernter: "Ueber die Häufigkeit, die Dauer und die meteorologischen Eigenschaften des Föhn in Innsbruck," *S. W. A.*, CIV., 2a, 1895, 427-461. The influence of the foehn upon all meteorological elements is fully discussed in this treatise.

floods of December 1-4, 1872, in Carinthia, there was also an exceptional occurrence of violent foehn winds, with very low relative humidity in the valleys of the northern Kalkalpen, between Vienna and Salzburg.[1]

It cannot be assumed that the warm, dry, southerly wind crosses the southern Alpine valleys at a considerable distance above the earth's surface, for this wind is not observed even on the crests of the Alps, where the rise of temperature is slight and the air remains damp. The unusually high temperature, and the low relative humidity, are confined to the valleys in which the foehn blows, and are less and less marked as the distance from the mountains increases. The following data show the average weather conditions which prevailed simultaneously on the northern and the southern sides of the Alps, during twenty winter days on which there were foehn winds.

WEATHER CONDITIONS AT MILAN, BLUDENZ, AND STUTTGART DURING FOEHN WINDS.

Station.	Temperature.			Relative Humidity (%).			Weather.
	A.M.	P.M.	Eve.	A.M.	P.M.	Eve.	
Milan, -	° 3·2	° 5·1	° 3·9	96	93	96	Rain on 16 days. Variable winds.
Bludenz, -	11·1	14·0	11·5	29	22	28	SE., 5-8. Foehn.
Stuttgart, -	3·4	8·8	5·0	84	72	81	Rain on 10 days. Variable winds.

The observations given in the following table show that the air of the foehn which blows in the northern Alpine valleys does not have its characteristic high temperature on the Alpine summits, but becomes warmed only when it blows down into the valleys.

WEATHER ALONG THE ST. GOTTHARD PASS DURING THE FOEHN OF JANUARY 31—FEBRUARY 1, 1869.

Station.	Altitude (m.)	Temperature.	Relative Hum.	Wind.
Bellinzona, -	229	° 3·0	80 %	N. (Rain).
San Vittore, -	268	2·5	85	S. & SW.
Airolo, - -	1172	0·9	—	N. & S.
St. Gotthard, -	2100	–4·5	—	S. 2-3.
Andermatt, -	1448	2·5	—	SW. 2.
Altdorf, - -	454	14·5	28	S. (Foehn).

[1] *Z.f.M.*, VIII., 1873, 10-11.

The temperature at Andermatt was the same as at San Vittore, on the southern side, 1200 m. lower down. On Mt. St. Gotthard itself, although there was a strong south wind, the temperature rose but little, while the rise at Altdorf was very marked. The southerly wind, therefore, did not gain its high temperature until it descended from a relative altitude of about 1700 m. During the protracted foehn period at the beginning of January, 1877 (Jan. 1 and 3-10), the mean temperature departure at the southern stations, Lugano and Castasegna, was only $+4{\cdot}3°$; on the St. Bernard, at 2478 meters, it was only $+3{\cdot}7°$; while in the "foehn valleys" it was $+11{\cdot}4°$ at Altdorf, and $+13{\cdot}3°$ at Altstätten. The departures again became less on the lowlands of Switzerland, but they were greater than on the southern side of the Alps. Thus at Zürich the departure was $+6{\cdot}9°$; and at Basel it was $+8{\cdot}0°$. These examples must suffice to show that the foehn attains its high temperature, as well as its dryness, on its descent from the summits of the Alps on their northern side, and that these characteristics are not imported from a distance farther south.

Cause of the high temperatures and dryness of the foehn.—It is clear that a mass of air which warms so rapidly during its descent must become relatively very dry. Assuming, for example, that the air on January 31 and February 1, 1869, was saturated with water vapour at the altitude of the St. Gotthard, every cubic meter of air at the temperature of $-4{\cdot}5°$ would have contained $3{\cdot}5$ grams of water vapour. On reaching Altdorf, with a temperature of $14{\cdot}5°$, this air could have contained $12{\cdot}4$ grams of water vapour, and the relative humidity of this south wind would therefore have become 28 per cent.[1]

The high temperature of the foehn is explained by the law which has already been discussed, viz., that a mass of air, descending to levels where the pressure is greater, warms at the rate of 1° in every 100 m. This is also the law of the vertical increase of temperature in foehn winds. The observations made at Swiss stations between 300 and 2700 m. above sea-level show that the increase in temperature from the summits of the Alps down into the "foehn valleys" is at the rate of $0{\cdot}97°$ in every 100 m.; while, on the other side of the mountains (on the southern side during a southerly foehn), the vertical decrease of

[1] By taking into account the change of volume resulting from the change in pressure and in temperature (19° in this case), we should obtain a more accurate result in the folliowng way: A cubic meter of air from the altitude of the St. Gotthard reaches Altdorf with a volume of $\frac{58}{72}(1+0{\cdot}00367\times 19°) = 0{\cdot}86$, which, at a temperature of $14{\cdot}5°$, can at a maximum contain $10{\cdot}6$ grams of water vapour. Hence the relative humidity becomes 33 per cent.

temperature is only 0·44° per 100 m., which is the average value for winter.

Weather conditions which give rise to foehn winds.—The European daily weather maps, which are constructed on the basis of telegraphic reports concerning the weather conditions all over the country, have shown why the air occasionally rushes down from the crests of the Alps into the valleys at their base, and thus gives rise to a foehn.[1]

This phenomenon is connected with the advance of barometric depressions or storm centres from the Atlantic Ocean toward western Europe. When a barometric depression is central west, or northwest, of the Alps, on a line between the Bay of Biscay and Ireland, the air flows across the Alpine foreland, as a southeasterly or southerly wind, toward the region of lowest pressure. The air within the Alpine valleys is likewise drawn toward this same region. As the wall of the Alps prevents any direct supply of air from coming in from the south, the air must come down from aloft, from the summits of the Alps, to replace that which has been removed from the valleys. Thus the foehn results. The stormy character of the foehn gusts in the valleys, and their remarkably local occurrence and great irregularity as a whole, may be explained as the result of this mode of origin, combined with the influence of the complex topography upon the movement of the air. On the level country away from the mountains, the air which flows toward the barometric depression is not interfered with by irregularities of the surface, and its movement is horizontal. Here the violence of the foehn, as well as its high temperature and low relative humidity, are lacking. Here the warm, dry air, coming out of the valleys, soon becomes a part of the prevailing atmospheric movements. As the cyclonic depression and the general southwesterly storm approach from the west, the southerly storm wind occasionally appears considerably sooner in the northern Alpine foreland than does the foehn in the Alpine valleys.

The air on the southern side of the Alps remains calm for a long time after the foehn has begun to blow on the northern side, because the barrier of the Alps prevents a supply of air from coming in from the south in the lower strata of the atmosphere. The pressure falls much less rapidly on the southern than on the northern side, and the

[1] The angle of descent of the air which comes down from the crest-line into the valleys is not nearly so great as is generally supposed. The angle of descent of a current of air flowing from the top of the Tödi to Auen, in the Glarna valley, is 9° ; from the crest-line of the Titlis range to Engelberg, it is about 12° ; and from Creux du Champ to Ormont, it is 25° (Wettstein).

temperature is still low, as is usually the case during winter anticyclones. The temperature decreases slowly with increasing altitude. A rise of temperature on the southern side does not occur until later, and is usually of less amount than that in the northern Alpine foreland, and hence very much less than that in the valleys where the foehn is blowing. Heavy rain almost always occurs with the rise in temperature on the southern side of the Alps.

The extent to which the ranges of the Alps operate to keep the air near sea-level from flowing toward the cyclonic depression to the northwest is shown by the great differences in pressure between the northern and southern sides of the Alps. On the average of seven days with foehn winds, the barometric gradient in 1° of latitude was 2·3 mm. between Basel and Altdorf; and 7·3 mm. between Altdorf and Lugano. The difference along the latter line was, therefore, three times as great as along the former. In individual cases, the difference may become 10-12 mm.

The southerly or southwesterly storm thus first appears on the western coast of Europe and advances towards Central Europe and the Alpine foreland. The air which descends from the summits of the Alps at first does not come from a great distance to the south. Its relatively high temperature is sufficiently explained as the result of the normal, slow, vertical decrease of temperature (shown by the relatively mild temperatures of the upper strata), and of the rapid warming of the air in its descent. In winter, the average rate of decrease of temperature with increase of altitude is 0·45° in 100 m. Indeed, under the weather conditions which precede the foehn, this rate is usually still slower. On the other hand, the descending air warms at the rate of exactly 0·99° in every 100 m. Hence the air which is coming down from the crests of the mountains increases in temperature 0·54° in every 100 m. In 2500 m., the rise in temperature is 13·5.° Hence the temperature may easily rise to 14° or 15°, in midwinter, without there being any necessity of postulating an importation of air from more southerly and warmer latitudes. After a time, to be sure, air is also brought in from the southern side of the Alps. This involves an ascent of the air on the southern slopes, and likewise a considerable condensation of water vapour in this mass of rising air. In consequence of this condensation, the ascending air cools but slowly, and the conditions which are necessary for the marked warming of the air on the northern side of the Alps remain unchanged, or are even accentuated. As the higher strata of the atmosphere are relatively much colder in summer, when the vertical temperature gradient is 0·6°-0·7°

in 100 m., it naturally follows, from the explanation already given, that the foehn cannot cause so great a rise of temperature in summer as in winter, when the vertical temperature gradient is only 0·3°-0·4°. Furthermore, the cyclonic depressions which come from the Atlantic are fewer in number and much weaker in summer than in winter. Hence the primary cause for the descent of the air from the crests of the Alps to the valleys is present less often, and to a less marked degree, during the warmer months.

Valleys which, by their trend or topography, prevent or interfere with the movement of the air toward the cyclonic depression in the northwest, also seem to be unfavourable to the occurrence of foehn winds; as, for example, are the upper portion of the canton of Valais; the valley of the Aar between Brienz and Thun, etc. On the other hand, the Rhine valley, and the valley of the Reuss at Altdorf, which open to the north, have very violent foehn winds, as has the lower Rhone valley, which turns to the northwest at Bex. When the cyclonic depression moves farther to the northeast or east, the wind veers from south and southwest to west and northwest, and the foehn is followed by a rapid fall of temperature and heavy rain, while the weather on the southern side of the Alps becomes fine.

Secondary depressions as related to the occurrence of foehn winds.—Erk, Billwiller, and Pernter have lately made a detailed study of the pressure conditions which produce foehn winds on the northern side of the Alps. The results of these studies show that in the rear of the larger barometric depressions, which usually pass by on the western or northwestern side of the Alps, there are developed local depressions, with their own inflowing spiral wind systems, along the northern border of the Alps and in the Alpine valleys themselves, and that these are the immediate cause of the foehn. This fact also explains the extraordinary local violence of the stormy foehn winds, and the frequent occurrence of these winds in the eastern Alps, especially at Innsbruck, in apparent disregard of the general distribution of pressure. In the case of the foehn of October 15-16, 1885, in the Bavarian mountains, which was not only an exceptionally violent one, but was also remarkable for other reasons, Erk has shown the existence, and has traced the progression, of a secondary local system of spiral winds in the northern Alpine foreland.[1]

Similarly, in the case of the foehn of January 13, 1895, Billwiller has been able to show that under the influence of a large barometric

[1] F. Erk: "Der Föhnsturm vom 15 u. 16 Oktober 1885, und seine Wirkungen im bayerischen Gebirge," *M.Z.*, III., 1886, 24-31.

depression on the coast of Ireland, a secondary depression formed over central and northeastern Switzerland, and local depressions formed in the "foehn valleys" of the Aar, Reuss, Linth, Upper Rhine, and Ill.[1] Probably all foehn storms occur under the influence of such local depressions, which are too faint to appear on the ordinary weather map.[2]

Foehn on the southern side of the Alps.—If the explanation of the foehn which has just been given is correct, and if this wind has no connection with the Sahara, we may expect that foehn winds will also occur on the southern side of the Alps when the pressure is high in the northwest and low in the southeast; *i.e.*, when the conditions of pressure distribution are the opposite of those above described as causing a foehn in Switzerland. As a matter of fact, the foehn is by no means unknown on the southern side of the Alps. When the distribution of pressure is as just stated (*i.e.*, high in the northwest and low in the southeast), warm north and northeast winds blow in the southern valleys of the Alps, and sometimes the north foehn is observed as far away as Milan. Yet the north foehn does not, on the average, attain nearly the violence of the south foehn, and similarly, the barometric depressions of the Mediterranean Sea are much weaker and less numerous than those of the Atlantic Ocean. Places which are well known to have frequent north foehns are Bellinzona, Lugano, Castasegna, the Lake of Como, Riva, Brixen, etc.[3]

The foehn occurs much less frequently in the eastern Alps, east of the longitude of Innsbruck, but it is by no means unknown on the southern side of the eastern Alps. Violent foehn storms occasionally occur in the valleys of the northern Alps, while at the same time the southern side of the Alps, as well as Carinthia, have heavy rains.[4] On the southern side of the Hohen Tauern there is a north foehn.[5]

[1] R. Billwiller: "Der Föhn vom 13 Januar, 1895, am Nordfuss der Alpen und die Bildung einer Theildepression daselbst," *M.Z.*, XII., 1895, 201-209.

[2] The theory advanced by Hébert is thus partly confirmed and explained, although in a sense somewhat different from that intended by the author himself. "Étude sur les grands Movements de l'Atmosphère et sur le Foehn et le Sirocco pendant l'Hiver 1876-77," *Atlas Météorologique*, VIII. (*Z.f.M.*, XIII., 1878, 317-320).

[3] Accounts of foehn winds at various stations in Europe will be found as follows:—*Z.f.M.*, III., 1868, 561-574; XIX., 1884, 89, 192; *M.Z.*, V., 1888, 175-180; VI., 1889, 192; VII., 1890, 228-230; X., 1893, 152-153.

[4] *Z.f.M.*, VIII., 1873, 10-11; XX., 1885, 515-516.

[5] The following references relate to foehn winds in the central mountains of Germany:—Vosges, *M.Z.*, XI., 1894, 143-147; Riesengebirge, *ibid.*, XII., 1895, 463-467; and R. Assmann: *Einfluss der Gebirge auf das Klima Mitteldeutschlands.*

Foehn winds in Greenland.—It follows from the theory that other mountains must also have their foehn winds, and the accounts concerning the geographical distribution of foehn winds show that this inference is fully borne out by observation. The most interesting case is that of the foehn on the western coast of Greenland, which was admirably described by Rink many years ago,[1] although not by him called a foehn wind. This is a very warm, dry, easterly or southeasterly wind, which comes across the ice-covered interior of Greenland, and blows down on to the fiords in stormy gusts. This wind raises the temperature on the average 12°-20° above the mean in winter, and about 11° in spring and autumn. Even as far north as Upernavik (latitude 72·5° N.), a foehn on November 24, 1875, brought a temperature of 10°, which was 25° above the mean. The foehn of November and the beginning of December, 1875, blew for 18-20 days over the whole of western Greenland, and caused an average rise of temperature above the mean of 8° in the south, and 15° in the north.

The Greenland foehn has been carefully studied by Hoffmeyer, who has shown that it is to be explained in precisely the same way as the Alpine foehn. While the foehn of November and December, 1875, above referred to, was blowing, there was a barometric depression south of Davis Straits, and high pressure over the North Atlantic Ocean near Iceland. This distribution of pressure caused southeast and east winds over Greenland, and these winds, blowing down on to the fiords from the elevated interior of Greenland, where the altitude above sea-level is at least 2000 m., necessarily gained the high temperatures to which reference has been made. The foehn of western Greenland cannot get its high temperature directly from the North Atlantic Ocean because the temperature of this wind in southern Greenland (latitude 62° N.) in winter is sometimes as high as that found over the Atlantic Ocean in the latitude of the Azores.[2] Kane, at Rensselaer Bay, in latitude 78° N., and Nares, at Floeberg Beach, in latitude 82·5° N., felt the warmth of the Greenland foehn.

On the eastern coast of Greenland, the foehn comes from the west and northwest. In Scoresby Sound, for example, the temperature on

[1] H. Rink: "Physikalische Beschreibung von Nord- und Südgrönland," *Zeitschr. Allgem. Erdk.*, II., 1854; N.S., III., 1857.

[2] N. Hoffmeyer: *Le Foehn du Groenland*, Copenhagen, 1877. Review of above in *Z.f.M.*, XIII., 1878, 65-70. More recent discussions of the Greenland foehn are the following:—*M.Z.*, VII., 1890, 109-115, 268-271; A. Paulsen: "Ueber die milden Winde im grönländischen Winter," *M.Z.*, VI., 1889, 241-249; H. Stade: *Verhandl. Gesells. für Erdk.*, Berlin, XX., 1893, 356-358.

January 10, 1892, rose from $-21{\cdot}2°$ at 4 A.M. to $+6{\cdot}0°$ at 8 A.M., with a west-northwest wind which blew from the ice-covered plateau.[1]

In Augmasalik (latitude 65° N.), in eastern Greenland, the foehn comes from north and northeast.[2]

Foehn winds in other countries.—Iceland also has its foehn, as Hoffmeyer has pointed out.[3] At Hermannstadt, in Hungary, the dry warm southerly wind is called the *Rotenturmwind.* Kutais has dry, warm east winds, which are most common in spring. They blow down from the Suram Mountains, and often injure the tobacco plantations. At Resht, on the southern shore of the Caspian Sea, a hot, dry wind occasionally blows down from the Elburz Mountains, as has been described in detail by Tholozan.[4] The stormy south winds at Gilan are extraordinarily dry and warm. They occur with the greatest violence at Kodum, 24 km. from Resht, and are most frequent in autumn; but they also occur in winter and spring when the highlands are covered with snow. These dry, warm, southerly winds occur throughout Gilan and in the western portion of Mazenderan.[5]

Trebizond also has its foehn.[6] On the Lake of Urmia, in Persia, there is occasionally a strong west wind which Perkins associates with the Arabian *samun.* It often blows for three days, and although it comes across the high snow-covered mountains of Kurdistan, it is so hot and dry that it scorches the vegetation.[7]

On the southern island of New Zealand the northwest storms, after causing very heavy rainfall on the western slopes of the New Zealand Alps, blow down on to the Canterbury Plains, on the eastern side of the mountains, as hot dry winds.

[1] *M.Z.*, X., 1893, 24-25.

[2] See *M.Z.*, VI., 1889, 378-381, and the explanation given by Hann; see also A. Woeikof: "Klima und Föhne der Dänemark-Insel, Scoresby-Sund," *M.Z.*, XVIII., 1901, 5-10.

[3] *Z.M.*, XIII., 1878, 146.

[4] See also G. Radde: "Reise nach Talysch, Arderbeidshan und zum Sawalan," *Pet. Mitt.*, XXVII., 1881, 51-52.

[5] J. D. Tholozan: "Sur les Vents du Nord de la Perse et sur le Foehn du Gilan," *Comptes Rendus*, C., 1885, 607-611, *Ann. Soc. Mét. de France*, 1885, 95.

[6] *Z.f.M.*, XIV., 1880, 325-328.

[7] Rev. J. Perkins, in a letter quoted by J. H. Coffin: *Winds of the Globe*, Washington, 1875, 444 (*Smithsonian Contributions to Knowledge*, No. 268). Perkins also describes the land and lake breezes and other local winds on the lake of Urmia.

Regarding the "nor'wester" of these plains, Murrough O'Brien writes as follows :—"The plains are about 100 miles (160 km.) long by 30 (48 km.) to 40 miles (64 km.) wide, reaching from the sea on the east to a range of mountains varying from eight to twelve thousand feet (2450 to 3650 m.) in height. Strong winds are very prevalent on the plains : that from the nor'west is the strongest and most furious that blows. At the foot of the hills and on the plains it is a very dry and often a hot wind, unaccompanied with rain—the sky is a peculiar deep, dull blue, and any clouds there may be seem not to move. On the tops of the ranges there rest heavy black clouds which, notwithstanding the furious wind, remain fixed. In the upper valleys very heavy showers accompany the nor'wester, the snow melts, and the rivers which rise in the glaciers and upper valleys are very suddenly freshed, rising from ten to twenty feet (3-6 m.) in a night. This wind frequently dies away at night, and begins again before midday. On the plains the rainfall is small, probably less than thirty inches (760 mm.). On the western slope of the backbone range, the climate is very wet. This nor'west wind blows continually all through the summer, and is stronger than any wind I have ever felt : in the river beds, dust and pebbles are blown along furiously, and even on the grassy plains it is often impossible to ride against the storm. . . . The wind in its greatest fury does not reach entirely across the plain, and is often confined to the lower front ranges, and some few miles to the east. I had on one occasion to harvest (working for a week) a field of corn by moonlight, being entirely prevented from working by day by this wind."[1]

In the valley of Lake Ohau the nor'westers are particularly violent at all seasons. They are most frequent from October to March, especially in February. They usually begin about 10 A.M., and are followed by rain until evening.[2]

The German international expedition to South Georgia had an opportunity to observe foehn winds on that island. Under the influence of these warm winds the maximum temperatures of the winter were almost as high as those of the summer.[3] Travellers among the Andes have noted the occurrence of a warm wind blowing

[1] M. O'Brien, in a letter quoted by S. Haughton : *Six Lectures on Physical Geography*, Dublin, 1880, 102 ; quoted by W. M. Davis, *Am. Met. Journ.*, III., 1886-87, 442-443 ; see also F. v. Hochstetter : *New Zealand, its Physical Geography, Geology, and Natural History* (transl. from German), Stuttgart, 1867 ; and J. v. Haast : *Geology of the Provinces of Canterbury and Westland, New Zealand*, Christchurch, 1879, 198-199.

[2] Alex. M'Kay : "On the Hot Winds of Canterbury" ; Cockburn-Hood : "Observations regarding the Hot Winds of Canterbury and Hawke Bay," *Trans. and Proc. New Zealand Inst.*, VII., 1874, 105-112. The last-named writer considers these winds as a continuation of the hot winds of Australia. When the northwest winds blow for several days vegetation withers, and even the leaves of succulent plants are scorched until they may be rubbed to dust like dry tinder. These winds have a very depressing effect upon human beings and upon animals.

[3] *Deutsche Polarstationen*, II., 339.

on the eastern side of these mountains, and have ascribed the high temperature of this wind to volcanic origin, because there seemed to be no other way of accounting for the warmth of a blast of air descending from snow-clad mountain peaks.[1] The foehn at Kanazawa, in Japan, has been described by Knipping.[2]

Chinook wind in North America.—Especial interest attaches to the so-called *chinook* wind which occurs east of the Rocky Mountains in North America, and which is a factor of great importance in the climate of a belt of country of considerable extent lying along the base of these mountains. The following description of this wind is given by Ballou:—[3]

"The extreme severity of the winters in certain parts of our north-western States among the Rocky Mountains and along their eastern base, is much tempered by the prevalence of a mild westerly wind, locally called the chinook. Its name is derived from that of the tribe of Chinook Indians, living near Puget Sound. It is said first to have been applied by the early Hudson Bay trappers and voyageurs, who, meeting the wind while travelling towards the Pacific coast, and finding it particularly strong and warm as they approached the lands of this particular tribe, called it the chinook wind.

"It is described as a soft, balmy wind, varying in velocity from a gentle breeze to a steady gale. Though its temperature rarely exceeds 10°, yet, coming as it does when one is accustomed to a low temperature, it seems warm by contrast with the preceding weather. The thermometer often rises from below $-17{\cdot}8°$ (0° F.) to $4{\cdot}5°$ or 7° in a few hours, and the maximum temperatures of the winter months in the Rocky Mountain region nearly always are coincident with the occurrence of a chinook.

"The sky is usually clear while the warm wind blows, though observers often note a few leaden-coloured clouds of a kind seen only during the chinook. These clouds are described as pancake-shaped, with peculiarly smooth, rounded edges, and stand apparently motionless, high above the mountain ranges.

[1] J. Miers: *Travels in Chile and La Plata*, London, 1826, I., 282; N. H. Bishop: *A Thousand Mile Walk across South America*, Boston, 3rd ed., 1874, 239. Both quoted by W. M. Davis, *Am. Met. Journ.*, III., 1886-87, 507-508.

[2] E. Knipping: *Der Föhn bei Kanawaza*, Yokohama, 1890 [*M.Z.*, VII., 1890 (88)-(89)].

[3] H. M. Ballou: "The Chinook Wind," *Am. Met. Journ.*, IX., 1892-93, 541-547, with bibliography. See also A. B. Coe: "How the Chinook came in 1896," *Mo. Weather Rev.*, XXIV., 1896, 413.

"The continuance of a chinook is as uncertain as its coming. It may last a few hours, or for several days. With a change of wind the temperature falls rapidly, and winter weather once more sets in.

"The chinook wind possesses to a remarkable degree the power of melting snow, for it is not only warm, but appears to be dry. Although a foot or more of snow may lie on the ground at the beginning of a chinook, it disappears within a very few hours, often seeming rather to evaporate than to melt. For this reason the chinook is most welcome to the cattlemen on the plains of Montana and Wyoming. In fact, without it, stock-raising would be almost impossible, as the dried grasses of the plains, on which the cattle subsist, would otherwise be buried the greater part of the winter. To a few, however, this wind, instead of being hailed with delight as a break in the cold of the winter, is a source of much discomfort. . . . 'Many nervous people feel prostrated by it, and tremble and fidget incessantly during a chinook. The faster the wind, the worse these symptoms are. . . .'

"The effect of the chinook on trees, especially those near the foot of the mountains, is far from beneficial, as sometimes, in early spring, under the influence of one of these winds, they are tempted unseasonably to start their sap and to open their leaf buds, only to be nipped in a few days by a cold wave."

It is natural that the cause of the unseasonable warmth of the chinook should have been commonly ascribed to the Kuro Siwo, or warm Japanese current of the Pacific Ocean, very much as the foehn in Switzerland has been popularly supposed to have its origin in the Sahara. The real explanation of the warmth and dryness of these winds is the same in both cases.

Of Alberta, in western Canada, M'Caul says: —[1]

"The grand characteristic of the climate as a whole, that on which the *weather* hinges, is the chinook wind. It blows from west to southwest, in varying degrees of strength. . . . In winter the wind is distinctly warm; in summer not so distinctly cool. Its approach is heralded by the massing of dark cumulus clouds about the mountain tops, and a distant wailing and rumbling from the passes of gorges. Its effect in winter is little short of miraculous. When a *real* chinook blows, the thermometer often rises in a few hours from 20° F. below (– 30 C.) to 40° F. above zero (+ 5° C.); the snow, which in the morning may have been a foot (0·3 m.) deep, disappears before night; everything is

[1] C. C. M'Caul: "South Alberta and the Climatic Effects of the Chinook Wind," *Am. Met. Journ.*, V., 1888-89, 145-159.

dripping; but before another night falls all the water is lapped up by the thirsty wind, and the prairie is so dry that a horse's hoofs hardly make an imprint."

Another description of the chinook in Canada has been given by Ingersoll,[1] as follows :—

"This wind is marvellous in its effects. To it is due the pleasing dryness of even the deepest gorges and nooks in the rocks in summer, while in winter it clears the plains for hundreds of miles away from the mountains of nearly all the snow—always scanty in amount—with amazing celerity. . . . Near the mountains only a few hours suffice to lick up all the snow, except from the gullies, into which it may have drifted to a great depth. Cattle and horses find the grass exposed, and resume their feeding. The cold has done them no harm, for there has been no wet snow or sleet."

It is owing to the chinook that snow seldom remains long on the ground on the plains along the foot of the Rocky Mountains in Canada. On the Kootenay Plains, cattle are left out all winter. Farther east, towards Winnipeg, it is colder. Edmonton, 480 km. east of the base of the Rocky Mountains, is still within reach of the chinook. The effects of these winds are noted from latitude 55° to 60° N. in Montana, Alberta, and in the Saskatchewan country, on the upper Peace and Liard Rivers. The exceptionally favourable climate of the Peace River and Saskatchewan country as compared with that of the same latitudes farther east, is due to the warm west winds from the Pacific Ocean, which, however, do not gain the characteristic high temperature and low relative humidity of the chinook until they reach the eastern side of the Rocky Mountains. At Isle à la Crosse, in latitude 56° N., potato plants are still green at the end of September, while in Manitoba they are exposed to frost as early as the middle of August. The fact that the wind does not become warm and dry until it reaches the eastern slopes of the mountains is shown by many interesting observations.

The westerly winds which blow down from the Rocky Mountains in latitudes lower than those which have been named above are dry and warm, but the conditions are no longer favourable for the development of strong foehn winds. Of Colorado Springs (lat. 39° N.) Loud says: "The excess of the westerly component (in the wind directions) here is due to the temporary prevalence of a wind quite characteristic of this region, and known locally (and somewhat facetiously) by the name of 'zephyr.' In the winter and spring it frequently blows with consider-

[1] E. Ingersoll, quoted by M'Caul, *loc. cit.*, p. 149.

able force and fury, usually from a point slightly north of west, and is characterised by marked warmth and dryness—qualities which suggest an origin similar to that of the Swiss foehn."[1]

The sirocco.—In the Latin countries of southern Europe, the name *sirocco* is locally used to designate a foehn wind. In general, the sirocco is a warm, and usually damp, southerly wind, whose characteristics are in striking contrast with those of the cold, dry, northerly wind. Thus the sirocco of Italy and of the Dalmatian coast is a damp, muggy, south or southeast wind. When, however, the southerly wind blows from the interior of an elevated district, and comes down from the crest of a mountain, or over the edge of a plateau, this wind is hot and dry—it is, in fact, a genuine foehn. Bridone noted, many years ago, that the hot sirocco of Sicily could not come from the African desert, because it would then be most violent on the south coast, whereas it actually occurs with its greatest velocity on the northern coast, especially at Palermo. Zona has more recently given a careful description of the well-marked sirocco of August 29, 1885. At Palermo, the temperature, at 1 P.M., rose to 49·6°; and the relative humidity fell below 10 per cent.; while at the same time, the temperature and humidity in the eastern and the southern portions of the island were not abnormal. The temperature rose to 42°; and the relative humidity fell to 16 per cent. between Termini and Alcamo, in the vicinity of Palermo. The sirocco which prevailed at Palermo had therefore not come across the interior of Sicily,[2] and could not have come from Algeria and Tunis, because the temperature on the coasts there was hardly 30°.[3]

On the west coast of Messenia, between Pylos and Kyparissia, there is a *sirocco di Levante.* This foehn-like wind sets in from southeast in the morning, and shifts during the day to east, and even to north. It is a very hot, dry wind, which dries the leaves of the trees so that they fall to the ground.[4]

The sirocco of the Algerian coast owes its origin to the same causes. A warm, southerly wind becomes 10°-15° or more warmer, and also becomes drier, during its descent from the mountains to the coast.

[1] F. A. Loud: "The Diurnal Variation of Wind-Direction at Colorado Springs," *Am. Met. Journ.*, I., 1884-85, 353.

[2] It must, however, be conceded that the temperature of the air was raised somewhat during its passage over the warmed surface of the ground.

[3] *Archiv. des Sci. phys. et nat.* (Geneva), XIX., 1888, 275-277; *M.Z.*, V., 1888, 409-410.

[4] A. Philippson, *Verhand. Gesell. für Erdk.* (Berlin), XV., 1888, 315-316.

During the sirocco of June 20, 1874, at Algiers, which has been described by Sainte-Claire Deville,[1] the thermometer quickly rose, at the beginning of the south-southeast wind, to 38·8°, while the relative humidity fell below 15 per cent.

On the northern coast of Spain, there is a hot, dry, southerly wind, whose presence may often be detected in the daily telegraphic weather reports from Bilbao. During the sirocco of September 1, 1874, the temperature at Biarritz, on the coast, rose to 38°; and Piche, with a sling thermometer, obtained a reading of 38·5° on the dunes of St. Jean de Luz.[2] The relative humidity was below 37 per cent. as late as 4 P.M., during the prevalence of strong south and southeast winds caused by a cyclonic depression off Ireland.

On the northern (French) side of the Pyrenees, the warm, dry wind is known as the *vent d'Espagne.* This wind blows during the approach of the larger depressions from the southwest, and rain is sure to follow, with a shift of wind to southwest, west, and northwest. Thus, as Piche has pointed out, the following, apparently paradoxical, proverb is true for the vicinity of Pau: *Plus il fait sec, plus la pluie est proche* (The drier it is, the sooner comes the rain).

The foehn is known as the sirocco as far as Innsbruck. Occasionally, the dry, warm, northerly winds (north foehn) along the southern side of the Alps, in the district of the upper Italian lakes, are called *siroccos*, the popular belief being that, because of their high temperatures, they must be southerly winds which have been turned aside from their onward course by the Alps, and as it were reflected by these mountains.[3]

These examples of the geographical distribution of the foehn must suffice. With the increasing number of meteorological observations, the phenomena associated with foehn winds will be discovered in many other places. It is to be expected, however, that these phenomena will be limited to the temperate and the polar zones, which are visited by well-developed cyclonic depressions and storms. The author knows of no case of a real foehn wind within the tropics, although the occurrence of such a wind may be looked for when a tropical cyclone approaches a high mountain range.

The bora and the mistral.—The explanation of the characteristics of the foehn on the ground that the latter is a descending wind

[1] C. St.-Claire Deville: "Coup de Sirocco, éprouvé à Alger le 20 Juin 1874, et suivi sur une grande Partie de l'Algérie," *Comptes Rendus*, LXXIX., 1874, 278-284.

[2] A. Piche: *Le Coup de Sirocco du* 1 *Septembre*, 1874. Pau, 1876.

[3] J. Hann: "Der Scirocco der Südalpen," *Z.f.M.*, III., 1868, 561-574.

(*fallwind*) has been doubted by some persons because the *bora*, which is certainly a very marked example of a down-cast wind, is cold. The bora of the Istrian and Dalmatian coasts is a strong northeast wind, which rushes down from the crest of the mountains on to the sea as an icy-cold blast, and is often accompanied by cloudy weather. The bora at Novorossisk, which has been described by Baron Wrangel, is still colder.[1]

Why does not the bora become warmed during its descent, and why does it not reach sea-level with a high temperature, instead of being so cold? The contradiction in this case is only apparent. The air of the bora, as well as that of the foehn, is warmed during its descent from the mountain tops, but the initial temperature in the former case is so low that, notwithstanding this warming, the wind reaches the warm coasts of the Adriatic, or of the Black Sea, as a cold wind. The bora occurs only in those localities where the back-country is very cold as compared with the coast; where, in other words, the isotherms are crowded closely together. The difference in temperature between the coast and the cold, elevated interior is so great that, even with an increase of temperature of 1° in every 100 m. of descent, the bora still appears as a cold wind. Moreover, the bora is really not so cold as it seems to be, its temperature being seldom below freezing. The crest of the Waradáh, across which the bora of Novorossisk blows on its descent, is only about 600 m. in height, but, according to Wrangel, the vertical decrease of temperature is nearly 2° in every 100 m., and occasionally it is undoubtedly much more rapid. Similarly, the country inland from the Adriatic coasts is very cold in winter. At Gospic, the mean minimum temperature in January is $-16{\cdot}7°$; while it is $0{\cdot}2°$ at Lussinpiccolo. The absolute extremes are $-27{\cdot}1°$ and $-2{\cdot}0°$, and the difference in altitude is 560 m. Even when the cold air of the Karst plateau warms at the rate of 1° in every 100 m. of descent, it still reaches the coast and the islands as a cold wind.[2]

[1] F. Wrangel: "Die Ursachen der Bora in Noworossisk," *Repert. f. Met.*, V., No. 4, St. Petersburg, 1876 (*Z.f.M.*, XI., 1876, 238-240). For a description of the bora of the Karst, the following important article should be referred to: F. Seidel: "Bemerkungen über die Karstbora," *M.Z.*, VIII., 1891, 232-235.

[2] See W. Köppen: "J. Hann über den Föhn von Bludenz," *Z.f.M.*, XVII., 1882, 461-468, and "Franz A. Velschow und der Föhn," *M.Z.*, IX., 1892, 75-76. Further, W. Trabert: "Zur Theorie der Erwärmung herabsinkender Luft," *ibid.*, 141-143; Hugo Meyer: "Ueber Fallwinde," *Das Wetter*, IV., 1887, 241-246; R. Klein: "Der Nordföhn zu Tragöss.," *Zeitschr. deutsch. u. oesterr. Alpenver*, 1900, 61.

A similar condition exists in the case of the *mistral*, a stormy northwest wind which occasionally blows down from the Cevennes, in the department of Provence, in southern France. The difference in temperature between the cold plateaus of central France and the warm southern valley of the Rhone, and especially the Mediterranean coast from Montpellier to Toulon, is so great that even air which has been warmed during its descent from the mountains must still appear as a cold wind. It should, moreover, be remembered that the bora and the mistral usually blow only when the interior districts are under the influence of cold which has been imported from more northerly latitudes. The amount of cooling is therefore very considerable, and the difference of temperature, which even under ordinary circumstances is very large, becomes still further exaggerated.[1]

Both bora and mistral occur when a barometric depression central over the southern Adriatic, or the Gulf of Lyons, draws the air out from the interior; or when a rapid rise of the barometer over the interior markedly increases the usual pressure gradient to the south. The southern Adriatic often has a sirocco from the southeast while a bora is blowing at Trieste and Fiume on the northern side of the barometric depression. Under these conditions, the bora may also be accompanied by cloudy weather, and even by rain and snow. When a depression lies over central or northern Italy, the Dalmatian coast has a sirocco and warm weather, while a mistral, with low temperatures, prevails in Provence. For example, on December 2, 1886, a cyclonic depression (750 mm.) was central at Leghorn. Toulon had a temperature of 2·3° and Perpignan had 5°, with a mistral blowing. Lesina, on the other hand, had 13° with a sirocco from south-southeast, accompanied by rain; and Pola had 9°, with heavy rain. On the 21st of the same month, when there was a depression over northern Italy (750 mm.), Perpignan had 3°, and Toulon had −1°, with a mistral from northwest. At the same time, Pola and Trieste had 15°, Lesina 16°, and Rome and Naples 15°. At all the last-named stations, a sirocco was blowing.

The cold, descending currents of air in valleys at night also arise from a difference of temperature which the warming during descent cannot overcome. Furthermore, this descending air continues to lose heat on the way down, and the descent is too slow to cause any appreciable warming as in the case of the foehn.

[1] O. Dersch: "Ueber den Ursprung des Mistral," *Z.f.M.*, XVI., 1881, 52-57. See also Sonrel: *Ann. Soc. Met. de France*, XV., 1867, 45.

CHAPTER XX.

MOUNTAINS AS CLIMATIC BARRIERS.

Effect of mountains on cloudiness and relative humidity.—The effect of a range of mountains upon the characteristics of the different winds is well shown by the accompanying data for the amount of cloudiness with different wind directions in southern Norway, as determined by de Seue.[1]

CORRELATION OF WIND DIRECTION AND CLOUDINESS IN SOUTHERN NORWAY (0-10).

	N.	NE.	E.	SE.	S.	SW.	W.	NW.
West coast, - -	7·0	5·0	4·6*	6·0	7·2	8·3	**8·5**	7·9
East coast, - -	4·2	5·7	7·3	**7·8**	7·6	5·7	3·8	2·5*

On the west coast, the west winds bring the maximum cloudiness, while on the east coast, the east winds are the most cloudy. Thus in both cases the winds which blow toward the mountains are the most cloudy, while the winds which come across the mountains are the clearest and the driest. This is particularly well shown in the case of the northwest wind on the eastern side of the mountains. The same thing is true for the relative humidity, as may be seen in the following table:—

CORRELATION OF WIND DIRECTION AND RELATIVE HUMIDITY IN SOUTHERN NORWAY (PER CENT.).

	N.	NE.	E.	SE.	S.	SW.	W.	NW.
West coast, - -	80	77	74	71*	72	79	83	**83**
East coast, - -	75	79	82	**85**	86	80	72	66*

[1] C. de Seue : " Windrosen des südlichen Norwegens," Christiania, 1876 (*Z.f.M.*, XII., 1877, 189-192).

A similar effect to that here noted in the case of the mountains of southern Norway is also seen in the case of other mountain ranges which lie more or less at right angles to the direction of the prevailing winds, and hence act as climatic divides. It should further be noted that calms are more than twice as frequent on the eastern side of the Norwegian mountains as on the western side (19 per cent. : 8 per cent.). As is well known, the west winds are the strongest.

Mountains as wind-breaks and barriers.—The fact last mentioned suggests the important climatic functions of mountains as wind-breaks and as checks to the interchange of air between the two sides of a mountain range. To a limited extent this effect may be seen in the relatively small average wind movement in a valley as compared with an open plain.[1]

The more luxuriant vegetation, and especially the more vigorous tree growth of such valleys, are effects not only of the more favourable conditions as regards temperature and of the more abundant precipitation, but also of the decreased air movement and the decreased evaporation as compared with the open country. Extended plains are exposed to more constant and more violent winds than are valleys, and such winds are, to a certain extent, unfavourable to tree growth. One of the difficulties in re-foresting great plains is the injury done by strong winds to the young trees. When considerable groves of trees once exist, they protect themselves against storms.

The Alps as a climatic divide.—In extra-tropical latitudes, mountain ranges which trend east and west, or which have approximately this direction, protect their southern slopes against cold northerly winds, and thus constitute important climatic bounds. The Swiss Alps furnish an illustration of this. If we cross one of the passes, such as the Brenner, the Splügen, the St. Gotthard, or the Simplon, from north to south, we are transferred in a few hours from the climate of central Europe to that of Italy. The climatic transition in this case is much more marked than it is between the eastern and the western ends of the range of the Alps. In fact, in crossing these Swiss passes the climate changes with great suddenness. The reason for this is to be found in the protection against the cold, northerly winds which the massive barrier of the Alps affords for the valleys on its southern side. It has also been seen that this protection is not always a negative one, so to speak, but is also positive in the sense that the cold wind, even

[1] Some narrow valleys which furnish a means of communication between larger valleys are, however, very windy, because the currents of air in them are accelerated, just as in the case of water in the narrower portions of river channels.

when it does descend across the mountains, becomes warmer during this descent. This beneficial influence of the Alpine range is most strikingly seen in the protected southern valleys themselves, the plain of northern Italy being colder and much more exposed than are these valleys. Villa Carlotta, on the Lake of Como, is 2·4° warmer in winter than Milan, and its minimum temperatures are 5° higher than those at Milan in winter. Riva is 1·5° warmer than Milan in winter. The mean annual minimum is −5·0° at Riva, and −8·0° at Milan. Even Bozen, which is 1° farther north and 110 m. higher, and lies at the foot of the Brenner, has precisely the same winter temperature as Milan. The mean annual minimum at Bozen is −7·7°, which is higher than that at Milan. These southern Alpine valleys have therefore been well named, "the fence of the garden of Europe."

When Dove says that the Alps have a cooling effect upon the plain of northern Italy in winter,[1] his meaning is easily misunderstood. The low winter temperatures of the plain of northern Italy are due, as has previously been suggested, to the fact that there is also a high enclosing mountain ridge on the south and west, which keeps out the warm winds from these directions in winter, while it allows free access to the colder winds from northeast and east. Of still greater importance is the further fact that the air which has been cooled as the result of radiation from the earth's surface has an excellent opportunity to accumulate over the plain. These cold masses of air stagnate over the lowlands, and are the cause of excessively low temperatures in winter,[2] the lowest temperatures always being found over the lowest portions of the plain in the axis of the valley of the Po. This characteristic is clearly seen even in the mean winter temperatures (1866-1880), as appears from the following table :—

MEAN WINTER TEMPERATURES (1866-1880) IN NORTHERN ITALY.

Station.	Latitude.	Altitude (meters).	Temperature.
	°		°
Milan, - - - -	45·5 N.	147	2·3
Brescia, - - - -	45·5 N.	172	2·7
Alessandria, - - -	44·9	98	1·3
Pavia, - - - -	45·2	98	2·2

[1] H. W. Dove : "Ueber den Einfluss der Alpen auf das Klima ihrer Umgebung," *Monatsber. Berliner Akad. der Wiss*, 1863, 96-114, and *Zeitschr. f. Allg. Erdk.* XV., 1863, 241.

[2] Compare the Engadine, Carinthia, the Lungau.

The Ligurian coast, the Riviera, is another example of the effects which protection against cold winds and exposure to warm winds, together with a favourable situation, have in producing a climate such as is normally found several degrees farther south.

The Carpathians and the Himalayas as climatic divides.—The climate of the plains of Hungary also illustrates the protection afforded by the enclosing Carpathian Mountains against the direct entrance of the cold from northern Europe. This is best seen in the greatly decreased monthly ranges of temperature, and even the mean and absolute winter minima also show this same effect. Thus, for example, Nyiregyhaza is about midway between Vienna and Czernowitz, but the mean monthly minima at the three stations are as follows:—

MEAN WINTER MINIMUM TEMPERATURES AT CZERNOWITZ, NYIREGYHAZA AND VIENNA.

	Czernowitz.	Nyiregyhaza.	Vienna.
Monthly Minima, - - -	° −16·5	° −12·5	° −10·5
Difference, - - - -	4·0°		2·0°

The most striking illustration of the protection afforded by a high mountain range against the advance of the winter cold of continental interiors into lower latitudes, is found in the winter temperatures of northern India, when these are compared with those of southern China, in the same latitudes.

WINTER TEMPERATURES OF NORTHERN INDIA AND OF CHINA.

Station.	Latitude.	Temperature.	Station.	Latitude.	Temperature.
Canton, -	23° 12′ N.	° 12·5	Shanghai,	31° 12′	° 3·9
Macao, -	22° 11′ N.	15·4	Multan,[1] -	31° 10′	13·9
Calcutta, -	22° 33′ N.	20·9	Lahore,[1] -	31° 34′	14·0

The winter at latitude 31°, in northern India, is thus seen to be 10° warmer than that on the coast of southern China, in the same latitude,

[1] The altitude of Multan above sea-level is 128 m., and that of Lahore, 223 m. The mean temperatures in the table have been reduced to sea-level, the rate of increase of temperature from the altitude of these stations to sea-level being taken as 0·4° per 100 m.

2 A

and at latitude 22·5° N., the Indian winter is 7° warmer than that on the Chinese coast. The difference would be still greater if the comparison could be made with stations in the interior, instead of on the coast of China. The reason for this difference is found in the fact that China is exposed to the cold northwest winds from the interior of Asia, while the massive mountain barrier of the Himalayas wholly prevents an interchange of air between the interior and northern India. Northern India has the winter climate which is appropriate to its latitude, the cooling effect of winds coming from higher latitudes being prevented.

Mountains in North America as climatic barriers.—If there were, in the United States, a lofty mountain range trending east and west, instead of the Rocky Mountain ranges which run more or less parallel with the meridians, the States lying south of such a range would be protected against cold winds from the north. Under existing conditions, however, these cold winds blow all the way to the Gulf of Mexico without meeting any obstruction; and they occasionally cause extraordinarily low minimum temperatures even in the southern States. The mean temperatures in the southern States are, to be sure, higher than in eastern Asia on the same latitude; but they are much lower than those in northern India, as is shown by the following table:—

WINTER TEMPERATURES IN THE UNITED STATES IN LATITUDE 31°–32° N.

Stations.	Latitude.	Temperature.	Stations.	Latitude.	Temperature.
		°			°
Mt. Vernon,	31° 5′ N.	11·3	Natchez, -	31° 34′ N.	10·5
Savannah, -	32° 5′ N.	11·2	Ft. Jessup,	31° 35′ N.	10·7

As northern India, in the same latitudes, has a winter temperature of 14·0°, the southern United States are 3° colder. The difference is still more striking when the extremes of cold are compared.

Protection against cold afforded by mountains.—The extraordinarily low winter temperature of northeastern Siberia, as well as the prevailingly high pressure which accompanies it, and to which attention was first directed by Woeikof, are undoubtedly chiefly due to the fact that air which has been cooled by radiation is prevented from flowing away by high mountain ranges on the south, and especially on the east, toward the neighbouring Sea of Okhotsk. These mountains also prevent an interchange of air between their opposite sides. In higher latitudes, where the ground is snow-covered in winter, such a stagnation of the

cold air always means an increase of the winter cold. A high mountain range, unbroken by deep ravines and valleys, protects the country in its lee very effectively against the invasion of winds from a cold interior district of this kind behind the mountains. Under these conditions, the lowest temperatures usually occur along the valley bottoms; in any case, there is certainly no decrease of surface temperature with increase of altitude. Hence the amount of warming which the air from the cold back-country undergoes in crossing the mountains and descending on to the lowland, or to the coast, is sufficient to maintain a considerable difference of temperature between the two sides of the mountains. If we imagine a "cold pole" of this kind, separated from a neighbouring warm coastal district by mountains 1500 or 2000 m. high, then the cold air, descending from these mountains, will be warmed 15° to 20°, and its temperature therefore becomes very much milder.

There are many examples of this beneficial effect of a mountain range. Even the course of the winter isotherms in eastern Asia and in northwestern North America shows how a high mountain range brings about a crowding of the isotherms. Another example is the case of Dalmatia, whose narrow coast is separated from a very cold interior by an almost unbroken mountain range. Exceptionally low winter temperatures occur over this interior, and the winds very often blow down as a bora from the crest of the mountains. Yet the cold of this off-shore wind is so much modified by its descent that, notwithstanding its bora, the Dalmatian coast has very moderate extremes of cold and very high mean winter temperatures. Thus we find a mean annual minimum temperature of $-20{\cdot}4°$ at Gospic (latitude 44·5° N., altitude 570 m.) in Croatia, near to, but shut off from, the Dalmatian coast by the Velebich.[1] This is almost the same as the mean minimum temperature at Cracow, and is evidently due to the stagnation of the cold winter air in the valley of the Lika. Very low winter temperatures undoubtedly also occur over the whole interior of Dalmatia. Even at Janina (latitude 39·8° N.) the mean minimum is $-8{\cdot}0°$; and in January, 1869, the minimum was $-17{\cdot}8°$. The minima on the coast are, however, very moderate. Thus Fiume has $-4{\cdot}4°$, with an absolute minimum of $-9{\cdot}0°$, in January, 1869; Lesina has $-1{\cdot}6°$, with an absolute minimum of $-7{\cdot}2°$; and Ragusa has $-0{\cdot}9°$, with an absolute minimum of $-6{\cdot}0°$. The intensity of the cold is modified by the descent of the cold air. Thus a mountain range is a most effective protection against invasions of cold from a neighbouring cold district, and makes it possible for decided extremes of temperature to exist in close proximity.

[1] In January, 1893, the temperature actually fell to $-30{\cdot}1°$.

If, however, a mountain range of this kind is crossed by transverse valleys through which the cold air may drain away, the districts at the mouths of these valleys are abnormally cold. Thus Woeikof attributes the excessively low winter temperature of Vladivostok (January mean, $-15{\cdot}2^\circ$, at latitude 43° 9′ N.) to the fact that, at this point, the top of the divide, across which the cold air from the interior has to pass, is only 180 m. high. To the eastward, the mountains become higher again, and the coast is warmer. To the north comes the wide gate at the mouth of the Amur, and here lies Nikolaiewsk, with a January temperature of $-24{\cdot}5^\circ$. Still farther north the mountains again separate the cold interior from the coast, and Ayan properly has a temperature of but $-20{\cdot}1^\circ$ in January, although its latitude is 3° more northerly than that of Nikolaiewsk.[1]

In summer, on the other hand, mountain ranges protect the interior against the entrance of cold winds from the ocean, and thus very high summer temperatures may be reached close to the coast. Striking examples of this condition are found in northwestern North America, where the rainy coastal district of British Columbia, with its cool summers, is separated by the mountains from the neighbouring interior, which lies very close by, and has hot summers. The isotherms are much crowded in this region, and high summer temperatures extend far to the northward in the lee of the mountains. Similarly, the cool air of the Sea of Okhotsk is prevented by the coast range of mountains from lowering the high summer temperatures of eastern Siberia, although the winds blow on-shore during this season. The contrast between the weather conditions on the two sides of the Aldan Mountains, in summer, is said by Erman and Middendorf to be a very striking one. On the eastern side hangs a chilly fog, through which the sun can rarely be seen; on the western side there is hot, bright summer weather. In much lower latitudes a similar dividing wall is formed by the Coast Range of California, which separates the cool, damp summer of the coast from the hot summer of the broad valley of the Sacramento and the San Joaquin rivers. In this Californian case there is one of the greatest contrasts in temperature which exists within a limited distance. At Monterey, for example, in latitude 36·5° N., the July isotherm of 16° runs along the coast, while hardly three degrees and a half to the eastward, beyond the Coast Range, we find the isotherm of 34°. This means an increase of temperature of nearly $5\frac{1}{2}^\circ$ in one degree of longitude, or 6° in 100 km.

[1] A. Woeikof: "Die Vertheilung der Wärme in Ostasien," *Z.f.M.*, XIII., 1878, 209-218.

In Scotland, the shores of Moray Firth, Nairnshire, and the southern portion of Sutherland, owe their favourable climatic conditions to the fact that they are on the lee side of some extended and rather high mountain ranges. In crossing these mountains the prevailing westerly winds become dry and warm, *i.e.*, gain, to a limited extent, the characteristics of foehn winds. The air is much drier, and the sky much less cloudy, on the eastern side of these mountains than on the western, and for this reason certain crops ripen at latitude 58° N., in the valley of the Shin in Sutherlandshire, though they never ripen in Argyllshire, 2° farther south (Scott). In northern Sweden, also, the west winds are warm, dry, and clear, because they come across the Norwegian mountains. In the northern hemisphere, where the southerly winds are the warm ones, the northern sides of mountain ranges have more extreme changes of temperature than the southern sides. The southerly winds are warmer, drier, and clearer on the north side than on the south. When a change of weather comes, the cold, northerly winds blow with undiminished intensity on the north side; but they reach the south side somewhat warmer, and also drier and clearer. Hence the change of temperature is greater on the north than on the south.

During a spell of unusually warm weather in March, 1896, and also during the change which followed it, the temperatures at Vienna and Graz were as follows:—

TEMPERATURES AT VIENNA AND GRAZ IN MARCH AND APRIL, 1896.

MARCH 21-26.—CLEAR WEATHER WITH SOUTHERLY WINDS.

	7 A.M.	Mean Maximum	Mean Minimum.	Mean.
	°	°	°	°
Vienna (North Side), -	6·0	19·9	5·3	12·6
Graz (South Side), -	3·5	17·6	2·4	10·0

MARCH 27–APRIL 2.—CLOUDY AND COOL, WITH WESTERLY WINDS.

	7 A.M.	Mean Maximum	Mean Minimum.	Mean.
	°	°	°	°
Vienna (North Side), -	3·1	10·0	2·7	6·3
Graz (South Side), -	4·1	12·2	2·4	7·3

Under the influence of the southerly winds it was cooler, and after the change to west and northwest winds it was warmer, at Graz than at Vienna. The difference in the daily means before and after the change to cooler weather was 6·3° at Vienna, and only 2·7° at Graz. The mean reduction of the after-

noon temperatures caused by the change was 9·9° at Vienna, and only 5·4° at Graz.[1]

Assmann has expressed these conditions as follows: In consequence of the foehn-like descent of the air which has become dry on the southern sides, the northern sides of mountains have higher temperatures, less cloudiness, and increased insolation. Mountains considerably increase the ranges of temperature over the lowlands lying to leeward, and therefore give the latter a more continental climate. The windward sides of the mountains, together with their neighbouring foreland, have a somewhat more temperate climate, while the lee sides, to a considerable distance, have a climate of somewhat greater extremes.[2]

[1] The temperatures at Graz are reduced to the altitude of Vienna by adding 0·7°.

[2] R. Assmann: *Der Einfluss der Gebirge auf das Klima von Mitteldeutschland*, Stuttgart, 1886, 57, 72.

SECTION III.—CHANGES OF CLIMATE.

CHAPTER XXI.

GEOLOGICAL AND SECULAR CHANGES OF CLIMATE.

Evidence of geological changes of climate.—The fact that great climatic changes have taken place on the earth's surface is proved by the geographical distribution of the fossil remains of animals and plants, as well as by other records of the earth's history. But besides those remote stages in the history of the development of our globe, changes are taking place before our eyes, which make us realise that the distribution of the climatic elements is not absolutely constant. There are at least certain fluctuations about a common mean, even if no progressive changes in a given direction take place.

Of the geological evidences of a change of climate the most striking is the fossil Tertiary flora of northern Greenland, Spitzbergen, and Alaska. This bears witness to the fact that during Tertiary time there was, between latitudes 70° and 80° N., an indigenous flora which points to the existence in those regions of a climate similar to that now found in northern Italy. The lignite beds of Discovery Bay, Grinnell Land (latitude 81° 44′ N.), in particular, show that the forests of this high latitude, in Tertiary time, consisted of taxodium (distichum), poplars, elms, lindens, firs, elders, hazel-nut trees, willows, and birches, together with water-lilies and irises. Judging by the present geographical distribution of these plants, it may be concluded that the mean July temperature in the northernmost portion of Greenland was 17° – 18° in those days, whereas it is now 2° – 3°; that the January temperature probably did not fall below – 6°, whereas it now reaches – 35° to – 40°; and that the mean annual temperature was at least 5° – 6°, the present mean annual being – 18° to – 20°. The

Miocene flora of Spitzbergen (latitude 78° N.) points to a mean annual temperature of about 11°, and that of Disco Bay, in Greenland (lat. 70° N.), to a mean of about 13°. In short, at the present time we must travel 20° – 30° southward in order to find the mean temperatures that probably prevailed in Greenland and Spitzbergen during Tertiary time.

In contrast to these facts which indicate a warmer climate, we have evidences of the so-called Glacial period, which occurred at a much later date in the earth's history, and during which an ice sheet covered northern Europe nearly as far south as latitude 51° N., and North America as far as latitude 40° N.; while the snow-line in the Alps reached at least 1300 m. nearer sea-level than at present. There were glaciers in the Erzgebirge and the Riesengebirge, as well as in the Black Forest and in the Vosges. The climatic snow-line on these mountains was then 1200 m. above sea-level. Glacial geologists are now of the opinion that the great Ice age did not consist simply of one period of glaciation, but was made up of a succession of glacial and interglacial periods, and thus had an oscillatory character. The succession and the extent of these diverse climatic changes in Spain has been carefully discussed by Penck.[1]

During Miocene time the climatic conditions of Spain were similar to those which are now found in Morocco. Not only were the isotherms then about 12° farther north than at present, but the whole wind system was correspondingly displaced. The northern limit of the trade wind belt must therefore also have been about 12° farther from the equator than it now is, and there must have been a dry period in Spain. The Miocene flora of central Europe (Oenningen), according to Heer, also points to a mean annual temperature of about 18°. During the Glacial period, on the contrary, the snow-line in central Spain was at least 1000 m. lower than at present, as was also the case in the mountains of central Germany. This leads to the conclusion that the temperature was 4·5° – 5° lower. The displacement of the climatic zones by more than 20° of latitude in Spain therefore closely corresponds with the conditions in central Europe.

Evidence of secular changes of climate.—Even at the present time there is evidence of changes of climate of smaller amount. This evidence is based on the fluctuations in the size of glaciers and in the extent of inland lakes. We have been living during a period of

[1] A. Penck: "Studien über das Klima Spaniens während der Jüngeren Tertiärperiode und der Diluvialperiode," *Zeitschr. Gesell. f. Erdk. zu Berlin*, XXIX., 1894, 109-141.

marked retreat of the Alpine glaciers, although this great retreat has now ceased, and is being replaced locally by movements in the opposite direction.[1]

The lakes of eastern Turkestan are at present rapidly drying up. The level of the Lake of Balkash is said to fall 1 m. in 14-15 years, and Lake Alakul is turning more and more into a salt deposit. In his work on Turkestan, Muschketoff gives many illustrations of the progressive desiccation and change in the climatic conditions of that country.[2] Rossikoff also gives evidences of the drying of the lakes on the northern side of the Caucasus,[3] where many steppe-lakes have already wholly disappeared. Brückner has given a general review of the conditions of increasing desiccation in western Siberia,[4] and Sieger has summarised the fluctuations of the water-levels of lakes.[5]

Evidence of changes of climate in the interior basin of the United States.—The results of investigations concerning the history of the fossil or extinct lakes which lay west of the Rocky Mountains in the United States are of great interest in establishing the fact of repeated changes of climate in the past. These lakes were Lake Bonneville, of which the present Great Salt Lake is a last remaining trace, Lake Lahontan, in Nevada, and Lake Mono, in California, all of which belonged to the Quarternary.

The careful investigations carried on by Gilbert[6] led to the conclusion that there is evidence of five distinct stages in the history of Lake Bonneville, which are as follows:—First, a long period of drier climate and of low water, during which alluvial cones, now more or less buried by later lake deposits, were built along the bases of the mountain ranges

[1] On page 323, the altitude above sea-level of the lower ends of the glaciers in the Oetzthaler Alps was given as 2100 m. This figure has reference to about the year 1860. About the middle of the 1880's, the altitude was 2250 m., or 150 m. higher. See also: "Rapport de la Commission internationale de Glaciers," presented to the International Geological Congress, Paris, 1900; and Ch. Rabot: "Les Variations de Longueur des Glaciers," Geneva, 1900.

[2] *Nature*, XXXIV., 1886, 119; 237. [3] *Nature*, XLIX., 1893-94, 515.

[4] Gaea, 1887, 187. See also *Pet. Mitt.*, XXXIX., 1886, Litteraturber. 76 (on the desiccation in the Aral-Caspian depression) and *Scot. Geogr. Mag.*, V., 1889, 327.

[5] R. Sieger: *Mitt. K.K. Geogr. Gesell.*, Vienna, 1888, 95, 159, 390, 418; *Globus*, LXII., 312; LXV., 73; *Zeitschr. Gesell. f. Erdk.* (Berlin), XXVIII., 1893, 478-481.

[6] G. K. Gilbert: "Lake Bonneville," *U.S. Geol. Survey, Monograph I.*, 1890, summarised by R. DeC. Ward: "The Climatic History of Lake Bonneville," *Am. Met. Journ.*, VIII., 1891 92, 164-170.

enclosing the lake. Second, a period of moister climate, and of high water which, however, did not result in an overflow. Third, a period of perhaps complete desiccation, as is shown by the unconformity between certain deposits on the lake bottom, and the difference in their character, the yellow clay of the first high-water epoch being overlaid by masses of gravel and boulders which are evidently not of subaqueous deposition. This has been called the inter-Bonneville time of low water. Fourth, a period of high water, which resulted in an overflow of the lake at the northeast corner of the basin, which was the lowest point that could be found. This was at a height about 300 m. above the level of the present Great Salt Lake, and here a channel was cut to a depth of about 100 m., first through alluvium and then through solid rock. Fifth, the present period of desiccation has resulted in the practical extinction of the old lake, which had an extent of about 550 km. in length and 200 km. in breadth. The periods of high water Gilbert believes to have been coincident with two epochs of maximum glaciation.

The study of the former great lakes, Lahontan and Mono, in Nevada and California, led Russell to similar conclusions.[1]

The causes of the fluctuations in the levels of these lakes do not appear to have been elevations or depressions of the land, but climatic changes. The lakes began to disappear when evaporation exceeded the water supply, and this was a result of abnormally high temperatures, and a long continuance of summer weather conditions. The lakes began to fill when the water supply was in excess of evaporation, and this corresponds to a deficiency of temperature, *i.e.*, a continuance of winter weather conditions. It is not necessary to assume any special excess of rainfall, for a moderate degree of humidity, with a low temperature, and the consequent decrease of evaporation would suffice to fill the lake basins.

Great Salt Lake even now shows long periods of high and of low water. The level of the lake rose more than a meter from 1847 to 1854; and then fell a meter and a half till 1859 and 1860. In 1861 the lake began to rise again, and in 1868 had risen 3·5 m., and continued rising a few centimeters until 1873, when it again began to fall. A granite post which Professor Joseph Henry had set up as a gauge in 1874, on the southern shore of the lake, was so far out in the lake in 1877 that a new gauge had to be set up further inland.[2] For

[1] I. C. Russell: "Lake Lahontan," *U.S. Geol. Surv., Monograph XI.*

[2] J. W. Powell: *Report on the Lands of the Arid Region of the United States*, 2nd ed., Chap. IV., Washington, 1879.

40 years a lake gauge has been maintained on Great Salt Lake by the U.S. Geological Survey. The zero of the scale was arbitrarily placed at what was supposed to be 0·3 m. (1 ft.) below the lowest known level. On December 31, 1900, the level of the lake was 230 mm. (9 ins.) below the zero of the scale. A tabulation of the rainfall records for three stations on or near the lake, since 1863, shows that the last fifteen years constitute the driest period of equal consecutive duration since the establishment of the Weather Bureau stations. In the light of the available data, Murdoch believes that there is no indication of a change in the climate of the Great Salt Lake basin; that a return to normal precipitation is probable, and that the lake will then return to about the same level as it reached in the past.[1]

Gilbert is of the opinion that variations in precipitation are a true cause of lake fluctuations, but believes that the increasing diversion to cultivated fields, of waters which would otherwise flow unimpeded to the lake, is also an important factor in this complex problem.[2]

Topographical records of past climates.—The relation of climate and topography has recently been considered by Davis.[3]

The dependence of certain topographic forms upon the conditions of greater or less rainfall of cooler and warmer climates are clearly recognised, and the correlations thus gained from the study of existing conditions may enable us to infer the vanished climates of the past by means of their still-preserved topographic products. The effect of the introduction of a dry climate in a region where rainfall had formerly been in abundance, and of a wet climate in a region that had previously been arid, may be seen in the changes of stream grades; in the conditions of valley-cutting, and in the increase and overflow or the desiccation of the lakes.[4]

Suggested causes of changes of climate.—Phenomena such as those which have just been described naturally prompt us to seek the causes which may have brought about these changes of climate in the past, and

[1] L. H. Murdoch: "Relation of the Water Level of Great Salt Lake to the Precipitation," *Mo. Wea. Rev.*, XXIX., 1901, 22-23.

[2] G. K. Gilbert: "The Water Level of Great Salt Lake," *ibid.*, 23-24.

[3] W. M. Davis: "A Speculation in Topographical Climatology," *Am. Met. Journ.*, XII., 1895-96, 372-381.

[4] See also in this connection Th. Rucktäschel: "Ungleichseitigkeit der Thäler und Wirkung der vorherrschenden westlichen Regenwinde auf die Thalformen," *Pet. Mitt.*, XXV., 1889, 224-225 (*M.Z.*, VII., 1890, 34-35, 180-182).

also to ascertain the facts concerning any changes and periodicities of climate which may be in progress at the present time. Climatic changes may either be progressive in one direction, or they may recur periodically, oscillating within certain fixed limits. The former class includes the changes that depend upon the progressive cooling of the earth's interior, a process which results in the gradual disappearance of the influence of the earth's internal heat upon the climates on its surface. To this first class belong also such changes as depend upon the cooling of the sun itself, the result of which is a progressive diminution in the intensity of solar radiation. In seeking the cause of the changes of climate which recur periodically, and which belong to the second class, we must begin with an examination of the periodic variations of the elements of the earth's orbit in order to determine whether these variations can explain the phenomena in question.

It is not our present purpose to consider the effect which a higher temperature of the earth's interior may have had upon the temperatures of the circumpolar regions, or upon a more uniform distribution of terrestrial climates. Such an influence may easily be overestimated, as has been shown by Sartorius von Waltershausen.[1] The influence of the earth's internal heat must have decreased rapidly after the surface became rigid, on account of the poor conductivity of the solid crust, which must soon have made the surface temperatures independent of the high temperatures of the interior. On the other hand, it is not quite so easy to see clearly what influence the oceans and their currents had upon climate in this transition period. A large amount of water vapour, and possibly also a higher percentage of carbon dioxide in the earth's atmosphere, may then have considerably increased the supply of heat from the sun stored up at the bottom of the atmosphere, and may, at the same time, have decidedly checked the cooling of the higher latitudes in winter. A more powerful radiation from the sun and a greater extent of water surface probably combined to maintain a humid atmosphere during the earlier ages. This condition may be assumed to have lasted for a very long time, during which the higher internal temperature of the solid crust of the earth no longer had any considerable direct influence upon the temperature of the ground.

Theory of Dubois.—The relations between a decrease or a change in the intensity of solar radiation and variations in the climates of the earth's surface must also be passed by without discussion. This

[1] S. von Waltershausen: *Untersuchungen über die Klimate der Gegenwart und Vorwelt,* Harlem, 1865.

question has lately been very carefully and fully discussed by Dubois,[1] whose theories have been critically reviewed by Woeikof in a very instructive article.[2] To this review the reader is here referred.

In so far as the conclusions reached by Dubois regarding climatic changes, and the explanation of these changes, depend upon purely geological facts, we must forego discussion at this point. We therefore turn at once to the cosmic causes of terrestrial climatic changes which are periodic in character, and as to whose effects we can, to some extent, make a quantitative estimate. Such estimates are not possible in the case of Dubois' assumption that the sun, like a star, has already passed through certain stages, during which the amounts of heat emitted by it to the earth varied. In the first stage, that of a white star, solar radiation was more intense, and the earth was warmed more than at present. With the transition to the yellow stage there comes a rapid decrease in the amount of radiation emitted, and hence a cooling of the earth, which Dubois believes to have lasted from the beginning of the Tertiary to the Pleistocene. During the yellow stage, there are variations in the amount of radiation emitted; the star temporarily has a red colour, and then radiation decreases. According to Dubois, these periods of partial darkening of the sun are glacial periods, while the return of the yellow light corresponds to the more extended inter-glacial epochs, in one of which we are living. These assumptions do not contradict our present knowledge of astrophysics. It is, nevertheless, impossible to reach any definite conception of the influence of these hypothetical variations in solar radiation upon terrestrial climates. One thing, however, is certain, and this is that, in explaining the climates of the geological past, the sun cannot be considered to be an absolutely constant source of heat. Hence, such considerations as those which have been urged by Dubois are perfectly legitimate.

Changes in the obliquity of the ecliptic.—Let us next consider the sun as a constant source of heat, and see what variations in the amount of heat received by the earth from the sun are produced by the periodic variations in the different elements of the earth's orbit. The climatic changes which result from this cause may be estimated with some accuracy. The variations in the amount of radiation can, as a matter of fact, be accurately calculated, but the effect of these variations

[1] Eugen Dubois: *Die Klimate der geologischen Vergangenheit und ihre Beziehungen zur Entwickelungsgeschichte der Sonne*, Leipzig, 1893; *The Climates of the Geological Past*, London, 1895.

[2] A. Woeikof: "Geologische Klimate," *Pet. Mitt.*, XLI., 1895, 252-256.

upon changes of climate is a much more complex problem than at first thought appears.

Of the periodic changes in the elements of the earth's orbit, we need to consider the change in the obliquity of the ecliptic, the changes in eccentricity of the earth's orbit, and the unequal length of the seasons. The change in the length of the seasons results from the precession of the equinoxes, which gives first one hemisphere and then the other a shorter summer and a longer winter.

The changes in the obliquity of the ecliptic may, according to Laplace, amount, at a maximum, to 1° 22·5′ on both sides of the value 23° 28′. The obliquity of the ecliptic may therefore attain the extreme values of 22° 6′ and 24° 50′. An increase in the obliquity of the ecliptic corresponds to a decrease in the total amount of heat received on the equator, and to an increase in the amount received at the poles. There is, therefore, under these conditions, a more uniform distribution of temperature over the earth's surface. The reverse is true for a decrease in the obliquity of the ecliptic.

Meech has calculated the amounts of heat for each 10° of latitude in the 10,000th year before 1800, when the obliquity of the ecliptic was near a maximum, and amounted to 24° 43′. The eccentricity at that time (0·0187) was somewhat greater than at present (0·0168). The former value was used by Meech in his determination, but we shall not consider this at the moment. Meech finds the change in the amount of heating at the different degrees of latitude about 1850, expressed in thermal days, as follows:—[1]

CHANGE IN THE HEATING OF THE EARTH AT A MAXIMUM OBLIQUITY OF THE ECLIPTIC, EXPRESSED IN THERMAL DAYS.

Equator.	10°	20°	30°	40°	50°	60°	70°	80°	Pole.
−1·7	−1·6	−1·3	−1·0	−0·2	0·7	2·1	5·5	7·2	7·6

The pole gains 7·6 thermal days, or 5 per cent. of its present insolation, while the equator loses barely 0·5 per cent. The middle latitudes remain unchanged. The difference between the opposite seasons becomes somewhat accentuated. It is therefore clear that the changes in the obliquity of the ecliptic can cause no considerable climatic changes.

[1] Cf. page 100. The unit is the heating effect of solar radiation in a mean day at the equator, which receives 365¼ of these units during a year. The pole now receives 151·6.

The eccentricity of the earth's orbit may vary between 0·0694 and nearly 0. The quantity of heat which the whole earth receives from the sun increases somewhat when the eccentricity becomes greater, the increase being inversely proportional to $\sqrt{1-\epsilon^2}$, where ϵ represents the eccentricity.[1]

If the quantity of heat received at the present time, when the eccentricity is 0·0168, is taken as unity, then at the maximum eccentricity the earth receives from the sun a quantity of heat equal to 1·0024, which is less than 0·3 per cent. more. The difference is therefore a slight one. On the other hand, the influence which a great eccentricity has upon the differences in the intensity of insolation at perihelion and at aphelion is important, as has already been shown in the note on page 96.

While, at the present time, the intensity at perihelion is only one-fifteenth greater than at aphelion, this difference increases at a maximum eccentricity to nearly one-third. In the hemisphere which, under these conditions, has its winter in perihelion, the difference between the seasons is considerably decreased, because the low altitude of the sun above the horizon is partially made up for by the less distance of the earth from the sun. In summer, on the other hand, there is decreased insolation because the sun's distance is much greater, notwithstanding the fact that the sun's altitude is high. In the other hemisphere, whose winter comes in aphelion, the difference between the seasons is similarly increased.

We must, however, be careful not to overestimate the influence, as there is a tendency to do. At the greatest eccentricity the solar constant (of Langley) would be 3·53 at perihelion and 2·47 at aphelion, and the mean value during the extreme seasons would be about 3·26 and 2·73. According to Angot, latitude 50° N. now receives, with a coefficient of transmission of 0·7, 22·9 heat days during the winter half-year, and 95·6 during the summer half-year, the difference being 72·7.[2]

At a time of greatest eccentricity, when the northern hemisphere had its winter in perihelion as at present, these values would be 24·9 and

[1] The amount of heat received by the earth from the sun in the course of a year is expressed by the formula, $CT/a^2\sqrt{1-\epsilon^2}$, in which T represents the time occupied in making one revolution around the orbit; a is one-half of the major axis; and C is a constant. The time occupied in making one revolution and the half of the major axis remain constant; therefore the heat can only vary with ϵ.

[2] See page 127. The coefficient 0·7 is too large for the direct transmission of solar rays, but may represent their virtual heating power when the effect of diffuse sky reflection, and the influence of the winds in bringing upper layers of air, warmed by the sun, down to the surface, are included.

87·0 (difference 62·1). The winter half-year would thus gain two heat days, and the summer half-year would lose 8·6. At latitude 50° S., the conditions at the same time would be the following:—Summer half-year, 103·9 heat days; winter half-year, 20·8; difference, 83·1. This difference would thus be one-third greater than that in the northern hemisphere. The southern hemisphere winter would have four heat days (about 16 per cent.) less than the northern hemisphere winter at the same time, while the southern hemisphere summer would have 16·9 days more, or about 20 per cent. more than the northern hemisphere summer.

Whichever hemisphere has a perihelion winter therefore has a modified solar climate, and a smaller annual range of temperature; while the other hemisphere at the same time has an extreme solar climate, with a large annual range. With the existing eccentricity of the earth's orbit, which is 1/60 in round numbers, as has been previously seen (page 201), the unequal distribution of land and water in both hemispheres has so great an influence upon climate that these theoretical relations do not appear at all. Moreover, existing climate is actually entirely different from the climate which results directly from solar radiation. The northern hemisphere, which is also the land hemisphere, whose winter comes at perihelion, has an extreme climate; while the southern, or water hemisphere, has a very temperate climate. Hence it may well be supposed that an extreme eccentricity of the earth's orbit would be unable wholly to overcome the climatic differences which at present exist between the two hemispheres. It would simply diminish these differences when the northern hemisphere winter came at perihelion, as at present, and would increase them when the northern hemisphere winter came at aphelion. The unequal distribution of land and water in the two hemispheres is the most potent climatic factor.

Differences in the length of the seasons: precession of the equinoxes.—A great eccentricity of the earth's orbit has another important result in that it can bring about considerable differences in the lengths of the winter and the summer half-years. The term *winter half-year*, or more briefly, *winter*, means that period during which the hemisphere in question always has the sun below the celestial equator, or when, in other words, the sun's declination is negative. *Summer* denotes the period during which the sun is above the celestial equator. Winter and summer are therefore the periods between the equinoxes. As long as the orbit is eccentric the only time at which the lengths of the opposite seasons are the same is when the spring equinox coincides with

perihelion or with aphelion. Under these conditions winter and summer would each last exactly half a year. When these conditions do not exist, the seasons are unequal in length. The inequality is greatest when the eccentricity is at a maximum, and when, at the same time, the line which joins the positions of spring and autumn equinoxes cuts the line of the apsides, which joins perihelion and aphelion, at right angles. The former line then divides the earth's orbit into two parts of unequal length. The shorter of these two portions of the orbit contains perihelion, and over this portion the daily movement of the earth around the orbit is also more rapid, because the distance from the sun is less. The longer portion of the orbit contains aphelion, and the velocity of the earth's progression along its orbit is here less. Hence it follows that the distance from the spring equinox to the autumn equinox (*i.e.*, summer) cannot be passed over as rapidly as the distance from the autumn to the spring equinox (*i.e.*, winter). When one hemisphere has its winter at perihelion, as is now the case in the northern hemisphere, the winter is shorter than the summer; while at the same time the opposite condition exists in the other hemisphere. The difference in the length of the seasons in days, with a given eccentricity, ϵ, is $465 \times \epsilon$. At present, when $\epsilon = 0{\cdot}0168$, the difference is 7·8 days. This means that the sun is north of the equator 7·8 days longer than it is south of it. At a time of maximum eccentricity, according to Lagrange, this difference may become 35·5 days, or, in other words, the difference may amount to more than one month. If we follow Leverrier in assuming 0·0777 as a maximum, the extreme difference in the length of the seasons is 36·1 days. Using Stockwell's value of the maximum eccentricity, 0·0694, which results when the perturbing action of Neptune, neglected by the earlier computers, is also included, the extreme seasonal difference is 32·3 days.

The progressive change in the position of the equinoxes amounts to 50·26 seconds annually. In a period of about 25,800 years the position of the spring equinox would therefore have travelled completely around the orbit. As the major axis of the earth's orbit also has a progressive movement, but in the opposite direction, the spring equinox returns to its original position in a shorter time; or, in round numbers, in 21,000 years. This is, therefore, the whole length of the period in question at the present epoch. Whenever, at any given time, the spring equinox coincides with perihelion or with aphelion, the difference between the seasons is zero. The difference then increases for, roughly, 5000 years, until it reaches a maximum; and then it

decreases again in the next 5000 years. Thus, during about 10,000 years, one hemisphere has the long winter and the short summer, and during the next 10,000 years a similar condition prevails in the other hemisphere. As the periods during which the eccentricity of the earth's orbit remains at a maximum are very much longer than these 10,000-year periods, these inequalities in the lengths of the seasons always occur several times during one period of great eccentricity, and they affect the two hemispheres in opposite ways.[1]

Adhémar's theory.—Many theories concerning great climatic changes which have affected the two hemispheres in opposite ways, have been based on these differences in the lengths of the winter and the summer half-years which recur periodically, and which, as has been seen, occasionally become very considerable. The first of these theories was that of the French mathematician, Adhémar. According to this theory, there is so great an accumulation of ice around the pole of the hemisphere which is having its long winter that the earth's centre of gravity is displaced somewhat toward this hemisphere. This results in a partial displacement of the ocean waters and a flooding of this hemisphere, which still further increases the cooling. The southern hemisphere, which is now having the longer winter, apparently illustrates these consequences. The curious forms of the southern portions of the continents of the southern hemisphere exemplify the flooding of the land; as do the enormous extent of the oceans of the middle and the higher southern latitudes, and the great extension of the lower limits of snow and of glaciers.

Because of its deceptive character, Adhémar's theory found many adherents, and, in a different guise, it has also often been brought forward again in more recent years.[2]

There is, however, as yet no proof that masses of ice could accumulate, in the hemisphere which has the long winter, to such an extent as to displace the earth's centre of gravity enough to cause a partial transfer of the surface waters of the oceans. The possibility of the displacement of the earth's centre of gravity by means of masses of ice about the pole, to a sufficient extent to produce a difference of level of over 800 feet between the Arctic and Antarctic Oceans, the Antarctic having the higher level at present, has been shown by

[1] A fuller discussion of these matters will be found in the text-books of astronomy.

[2] Adhémar: *Les Révolutions de la Mer ; Déluges périodiques*, Paris, 1842.

Proctor,[1] who also attributes the smaller barometric pressure of the southern hemisphere to the same cause.

Schmick's theory.—Schmick has endeavoured to refer the periodic transfers of water, which he also supposes to have taken place, to another cause, a mathematical and physical explanation of which seems to be easier. The southern hemisphere now has its summer in perihelion, and therefore that portion of the earth's surface which is nearest the sun is in the southern hemisphere. Hence, according to Schmick's calculation, the solar tide in this hemisphere is increased by 4 cm., and great quantities of water are drawn over into the southern hemisphere every day. This water cannot flow back again, because, when this process ceases at aphelion, the sun's attraction is weaker. Furthermore, the transfer of these great volumes of water south of the equator somewhat displaces the earth's centre of gravity toward the southern hemisphere, and this is followed by a partial flooding of this hemisphere. In conjunction with these conditions, there is a greater cooling in this hemisphere; more water is changed into ice, and this ice remains there. These floods and this refrigeration occur in the hemisphere which has its summer in perihelion. By reason of the apparently sound physical tidal theory upon which it rests, Schmick's hypothesis was long recognised and supported by scientific men of high standing. The most complete refutation of Schmick's views concerning the possibility of such a transfer of water from one hemisphere to the other by means of the solar tide is contained in Zöppritz's discussion of the theory.[2]

Croll's theory.—The theory of great periodic changes of climate advocated by James Croll has become of much more importance than the preceding theories, and has also gained a firmer and more lasting foothold. Croll's theory likewise depends upon the inequality in the length of the seasons at a time of great eccentricity of the earth's orbit. Croll, however, showed great skill and ingenuity in starting

[1] Proctor and Ranyard: *Old and New Astronomy*, Arts. 1092-1095, London, 1895.

[2] J. H. Schmick: *Die Umsetzungen der Meere und die Eiszeiten der Halbkugeln der Erde: ihre Ursachen und Perioden*, Cologne, 1869; *Die neue Theorie periodischer säkulären Schwankungen des Seespiegels und gleichzeitiger Verschiebung der Wärmezonen auf der Nord- und Südhalbkugel*, Münster, 1872; *Das Flutphänomen, und sein Zusammenhang mit den säkulären Schwankungen des Seespiegels*, Leipzig, 1879; *Die Aralo-Kaspi-Niederung und ihre Befunde im Lichte der Lehre von den säkulären Schwankungen des Seespiegels und der Wärmezonen*, Leipzig, 1874. K. Zöppritz: Criticism of these articles in the *Göttingische Gelehrte Anzeigen*, 1878, 865-871.

with a sound climatological basis as the foundation of his hypothesis. In so doing, he was enabled to make most effective use, in support of his theory, of the great influence of warm ocean currents in tempering the climates of the higher latitudes. It thus became necessary for him to advocate most strenuously the wind theory of ocean currents, about which there was, in his time, still considerable dispute; and for this very reason, Croll's theory will always occupy a prominent place in the history of our science, even when that theory has lost the last supporter of the principal portion of its teaching. It should not be forgotten that climatology owes much to Croll's hypothesis for its suggestion of new points of view; for the emphasis which it has laid upon climatic factors before largely overlooked, and especially for the instructive discussions to which the controversy regarding the hypothesis has given rise. Croll's theory still has a good many adherents.[1] For these various reasons, this theory demands somewhat careful attention.[2]

Croll's views may be summarised briefly as follows, the statements being largely those used by their original author:[3]

When winter occurs at aphelion during a period of very great eccentricity of the earth's orbit, that season is very much longer and colder than at present. Snow then falls over the temperate zones, even in latitudes in which the precipitation is now altogether rain, and although the snowfall may not be very heavy, the snow remains on the ground without melting because the temperature is far below the freezing point. On the approach of spring and summer, the rising temperature at first still further increases the snowfall over the lands, because of the increased evaporation over the oceans. When, how-

[1] A. R. Wallace: *Island Life*, London, 1892. The author is an advocate of a very much modified Croll's theory. James Geikie: *The Great Ice Age*, 3rd ed., London, 1894. The author still holds the view that Croll's theory gives the most probable explanation of glacial periods.

[2] James Croll first discussed his theory in a series of articles in the *Philosophical Magazine* and the *Geological Magazine*; the first article having the title, "On the Physical Cause of the Change of Climate during Geological Epochs," *Philosoph. Mag.* (London), 4th Series, XXVIII., 1864, 121-137. These papers were later published together in book form, with the title, *Climate and Time in their Geological Relations: a Theory of Secular Changes of the Earth's Climate*, London, 1875. A second book had the title, *Discussions on Climate and Cosmology*, London, 1889.

[3] These quotations may also serve to illustrate Croll's manner of drawing conclusions. These are frequently very uncertain, and therefore cannot easily be refuted.

ever, the snow begins later to melt, a long time must elapse before it wholly disappears from the lowlands. It will remain on mountains of moderate height, and with the coming of autumn there is a renewed fall of snow. In the following year the process is repeated, with the difference that the snow-line then descends to a lower level than in the preceding year. The snow-line thus descends lower and lower, year after year, until finally all the more elevated portions of the land remain permanently snow-covered. The valleys are then occupied by glaciers, and perhaps one-half of Scotland, a large portion of England and Wales, and nearly all Norway, may be covered with ice. Then a new and important factor comes into play, which very greatly hastens the process of glaciation. This is the influence of a snow-cover on climate. The extended surfaces of snow and ice change the water vapour which is brought by the winds into snow. They also cool the air during the summer, and produce dense, permanent fogs, which shut out the sun's rays, and give rise to climatic conditions such, for example as those which now prevail in South Georgia. The melting of the snow is largely retarded by these conditions.

It is a mistake to assume, says Croll, that the perihelion summers of the glacial periods must be hot. No continent which is covered with snow and ice can have a hot summer, as is proved by the conditions in Greenland to-day. Even India, if it were covered with ice, would have a colder summer than England now has. Another circumstance of the very greatest importance also comes into play, and that is the mutual reaction of the physical conditions. The great eccentricity produces long, cold winters in one hemisphere. The cold produces a more extended snowfall. The snow-cover, in its turn, increases the cold; it cools the air and gives rise to more snowfalls. A third factor then comes into play; namely, the formation of clouds and fogs, which partially shut out insolation, weaken the effect of the sun's rays, and increase the accumulation of snow. Thus the cold increases the snowfall, and the snowfall increases the cold. In other words, the effect operates to increase the cause.

While snow and ice are accumulating in one hemisphere, they are decreasing in the other. This increases the trade winds in the cold hemisphere, and weakens them in the warmer hemisphere. The result will be that the warm water of the tropical oceans is driven over, more and more, into the middle latitudes of the warm hemisphere. If, for example, the northern were the cold hemisphere, with the long aphelion winter, the Gulf Stream would thus continue to decrease in volume, while the warm ocean currents of the southern hemisphere

would at the same time increase in strength. This diversion of the sources of heat from the higher latitudes of the northern hemisphere again favours the accumulation of snow and ice over that hemisphere, and thus the warm ocean currents are still further weakened.[1] Thus these two consequences mutually strengthen one another.

A similar process of mutual action and reaction also comes into play in the warmer hemisphere, but in the opposite sense. Here everything co-operates to raise the mean temperature, and to decrease the quantity of snow and ice in the temperate and the polar zones. All these forces are set in operation by a great eccentricity of the earth's orbit, coupled with an extreme perihelion position.

The foregoing paragraphs contain the essential points in Croll's theory of glacial periods. The inter-glacial epochs correspond to the periods during which one hemisphere has the perihelion winter and, coupled with that, the long summer. The periods of equality in the length of summer and of winter, in both hemispheres, are periods of transition.

Objections to Croll's theory.—It may at once be urged against Croll's theory that the known climatic conditions of both hemispheres at the present time decidedly contradict the assumption of any such marked change for the worse in the climate of the hemisphere which has the long aphelion winter. At present, the difference in the duration of the opposite seasons is eight days; yet this condition of things does not seem to lead to any definite consequences. It is, moreover, perfectly clear that the longer winter of the southern hemisphere is milder than that of the northern; that the equatorial limit of winter snowfall is, on the

[1] This is one of the chief arguments in Croll's theory. Croll has, in several of his publications, endeavoured to prove the extraordinary importance of the Gulf Stream, and of the warm ocean currents in general, as means of modifying the temperature of the higher latitudes. The great volume of the warm ocean currents in the northern hemisphere is, however, caused by the extension of the southeast trade winds into that hemisphere, which results in driving the warm surface waters of the whole equatorial zone over into the northern hemisphere, and thus helping to supply these warm currents. This extension of the southeast trade winds is, however, a consequence of the low temperature of the southern hemisphere, and this low temperature is a consequence of the longer winter of that hemisphere. When this latter condition is reversed, so as to be unfavourable to the northern hemisphere, then the northeast trade wind will play the part which is now played by the southeast trade wind. The northeast trade wind will then blow across into the southern hemisphere, and will transfer warm water from the northern into the southern hemisphere. This will result in the diversion of the present warm currents referred to in the text, and will cause a marked change for the worse in the climate of the higher latitudes.

average, to be found in higher latitudes in the southern hemisphere than in the northern.

It is in the highest degree improbable that, even if the difference in the length of the seasons does become four times as great, conditions should undergo so decided a change as that which is demanded by Croll's theory. Furthermore, the effect of the short but hot summer upon the hypothetical accumulation of snow cannot be considered as unimportant as it is regarded by Croll. An Indian sun which can melt a layer of ice 50 to 60 m. thick in a year, or 17 cm. thick in one day, would quickly dispose of a snow-cover which certainly could not attain any such thickness over night.

The greatest importance is attached by Croll to the diversion of the warm ocean currents, which is supposed to occur, at a time of great eccentricity of the earth's orbit, in the hemisphere which has the long severe winter. This postulate rests on a very weak foundation. In this connection John Ball notes that the case of the southern hemisphere, with its mild winter and weak temperature gradient between poles and equator, shows clearly enough that the strength of the trade winds, and their extension into the other hemisphere, can by no means depend upon a low temperature and a severe winter. Therefore Croll has no right to assume that a cold hemisphere, with severe winters, will have unusually strong trade winds, which will drive the warmer water over into the other hemisphere.[1]

Woeikof further rightly calls attention to the fact that the reason for the extension of the southeast trade wind into the northern hemisphere is not that the southern hemisphere is now having its winter in aphelion, but that the extent of the oceans of the southern hemisphere is so great. This makes the trade winds of the southern hemisphere steadier and stronger. Land surfaces, and even small clusters of islands, check the development of the trade winds by giving rise to monsoons and land and sea breezes, and thus interfering with the steadiness of the trades. When the northern hemisphere happens to have the longer winter, it by no means necessarily follows that the northeast trade winds will be so much stronger as to extend over into the southern hemisphere. The fact that so much warm water is at the present time driven over into the northern hemisphere is also largely due to the form of the continents within the tropics, and especially to the trend of the coast-line of South America north of Cape San Roque. The position of the belt of calms between the two trade wind belts

[1] John Ball: *Notes of a Naturalist in South America,* London, 1887. "Appendix B, Remarks on Mr. Croll's Theory," 393-406.

does not depend upon the severity of the winter in the higher latitudes, but is determined by the general distribution of temperature in the lower latitudes, *i.e.*, upon the position of the heat equator. It is clear that, notwithstanding the severe winter of the northern hemisphere, the heat equator and the belt of calms do not migrate into the southern hemisphere. The position of the belts would remain about the same as long as there was no change in the distribution of land and water. The higher temperature of the oceans of the northern hemisphere is not alone due to the extension of the southeast trade winds across the equator, and to the warm ocean drift which depends upon these winds. The oceans of the northern hemisphere, which are shallower and more enclosed than those of the southern, are also warmed to a greater degree. Furthermore, the distribution of land about the north pole almost altogether prevents any supply of cold polar water and of floating ice from coming into these oceans, as is the case in the southern hemisphere. The amount of cooling by the melting of polar ice is therefore very slight in the northern oceans, but is very considerable in the southern hemisphere. A long aphelion winter in the northern hemisphere would change these conditions but little. In view of all these facts, the diversion of the Gulf Stream and of the warm ocean currents in general, in consequence of more severe winters, must be regarded as a very fanciful hypothesis. Howarth also calls attention to the fact that the drift of warm water to the higher latitudes is due to the prevailing westerly winds north of the 30th parallel. These winds, however, are stronger, the greater the difference of temperature between pole and equator; they would therefore be more effective in the hemisphere which has the cold winter.

Another consequence of a more severe winter, and hence of a stronger temperature gradient, has been pointed out by Davis.[1] The greater activity of the general circulation in extra-tropical latitudes would provoke more active cyclonic processes, and this would cause a greater snowfall. Further, an increased severity of winter temperatures would carry the subtropical belt of winter rains farther towards the equator, and thus give more precipitation over the continents.

Finally, as regards the phenomena of glaciation, it is now recognised as an established fact that glaciers do not depend upon a long severe winter. On the contrary, such conditions are unfavourable to the development of large glaciers, as may be seen in districts of severe winter cold in the interior of the northern continents, especially

[1] W. M. Davis: "Note on Croll's Glacial Theory," *Trans. Edinb. Geol. Soc.*, VII., 1894, 77-80. Reprinted in *Am. Met. Journ.*, XI., 1894-95, 441-444.

in Siberia. The case of the coast and of the interior of Alaska, to which reference has been made on p. 323, is also very instructive. It is not a severe winter, but a cool summer, which favours the descent of glaciers to low altitudes. It is not the extreme hemisphere, with the cold aphelion winter and the hot perihelion summer, that offers the most favourable climatic conditions for a great extension of glaciers; but it is the temperate hemisphere, with a slight difference between winter and summer, as may be clearly seen in the case of the southern hemisphere at the present time, and also in the glaciers on the west coasts of North America and of northern Europe. The smaller the annual range of temperature, the further toward sea-level do the glaciers extend, and the higher is the mean annual temperature at their lower limits. These facts therefore wholly contradict Croll's theory, which maintains that the greatest extension of glaciers will occur in the hemisphere with a severe winter and a hot summer.[1]

Glaciers are phenomena which are associated with mountainous coasts and islands. They do not find climatic conditions suited to them in the interior of the larger land areas. The winter in the interior of these continents in the higher latitudes is too cold and too dry, and the summer is too hot. Glaciers need a moderate marine climate, with abundant precipitation; and these two conditions are found together on the western coasts of the continents, while the eastern coasts, with their colder winters and their more severe climates, are unfavourable to the development of glaciers.

The interior ice of Greenland, unless it be a formation which dates from a former period in the earth's history, probably owes its existence to the fact that a land mass lies between two relatively warm bodies of water. These waters are not frozen over in winter; and for this reason, as well as because of the neighbouring warm North Atlantic Ocean on the south, they are the principal tracks of cyclonic storms which cause frequent and abundant snowfalls over the Greenland plateau. The winter temperatures along both coasts of Greenland are much milder, and even the summer temperatures are higher, than they are in the Chukt Peninsula in eastern Asia, in the same latitude, and yet, notwithstanding its mountainous character, the latter district has no glaciers.

[1] It cannot be denied, however, that the southern hemisphere, now in its aphelion winter, possesses much the largest accumulation of polar ice, forming a veritable ice cap; but whether this is due to the existence of an Antarctic continent permanently glaciated, or to causes which are independent of topography, can only be determined by further geographical exploration.

We therefore reach the conclusion that Croll's theory as to the cause of the great changes in climate which have taken place in the two hemispheres, and as to the resulting periodic glacial epochs, contradicts our present knowledge of climates, and is therefore untenable.[1]

Ball's theory.—The eccentricity theory of glacial periods has recently again been taken up by the English astronomer, Sir Robert Ball.[2]

As it might very naturally be supposed that this book treats the old problem in a new way, it may be said at once that such is by no means the case. The statement upon which the whole "new" theory rests, and which is reiterated in many different ways, is as follows:—The ratio of the quantity of heat which a whole hemisphere receives from the sun in winter to that received in summer is expressed by 37 : 63, and is virtually independent of the changes in eccentricity.[3]

Ball considered this a new statement, but it had already been used by Wiener in his important discussion of the distribution of the intensity of solar heat over the earth's surface.[4] Starting with this constant relation, Ball considers the temperature conditions of both hemispheres at a time of great eccentricity on the basis of the

[1] Of the instructive discussions and refutations of Croll's theory, the following may be referred to:—S. Newcomb: "Review of Croll's Climate and Time, with especial Reference to the Physical Theories of Climate maintained therein," *Am. Journ. Sci.*, 3rd Series, XI., 1876, 263-273. A. Woeikof: "Examination of Dr. Croll's Hypothesis on Geological Climates," *Philosoph. Mag.*, London, 5th Series, XXI., 1886, 223-240. H. H. Howorth: "A Criticism of Dr. Croll's Theory of Alternate Glacial and Warm Periods in each Hemisphere, and of Interglacial Climates," *Mem. and Proc. Manchester Soc. Lit. and Philosoph.*, 4th Series, III., 1890, 65-111. W. N. Rice: "The Eccentricity Theory of the Glacial Period," *Science*, VIII., 1886, 188-189, 347.

In connection with the relations between glaciers and mean temperatures, the following articles may be referred to:—Frankland: "Physikalische Ursache der Eiszeit," *Pogg. Ann.*, CXXIII., 1864, 419. G. F. Becker: "On the Relations of Temperature to Glaciation," *Amer. Journ. Sci.*, 3rd Series, XXVI., 1883, 167-175; "The Influence of Convection on Glaciation," *ibid.*, XXVII., 1884, 473-476.

[2] Sir Robert S. Ball: *The Cause of an Ice Age*, London and New York, 1891.

[3] If the sun's heat received by the hemisphere in one year is expressed by 1, the amount received in winter by $1-a$, and that in summer by $1+a$, half the annual amplitude, a, is equal to

$$2 \sin \delta : \pi,$$

if δ represents the obliquity of the ecliptic. When $\delta = 23° \ 27' \ 30''$, $a = 0{\cdot}253$; $1-a = 0{\cdot}747$, and $1+a = 1{\cdot}253$. Hence winter : summer $= 0{\cdot}37 : 0{\cdot}63$, or still better, $= 3 : 5$. This relation, however, holds only for the whole hemisphere, and as may easily be seen, not for the different latitudes. It therefore has no special significance for glacial theories.

[4] C. Wiener: "Ueber die Stärke der Bestrahlung der Erde durch die Sonne in den verschiedenen Breiten und Jahreszeiten," *Z.f.M.*, XIV., 1879, 113-130.

following calculation :—If the difference between the seasons has increased to 35 days, then the winter in the hemisphere which has its winter in aphelion lasts 200 days, and the summer 165 days. If the annual quantity of heat is taken as 365 heat days, the winter, according to the above proportion (37 : 63), has 136 of these days and the summer 229. The quantity of heat received on these 136 heat days is, however, distributed over 200 days, and the winter day therefore receives, on the average, only 0·68 heat units; while the summer day receives 229 : 165, or 1·39. The difference between a summer day and a winter day is therefore 0·71 heat units. This corresponds to the glacial period in that hemisphere. When this same hemisphere then has the long summer, a summer day receives 229 : 200, *i.e.*, 1·14 units of heat, and a winter day receives 136 : 165, *i.e.*, 0·82 units. The difference between the summer day and the winter day is then only 0·32 units of heat. This corresponds to the interglacial epochs. This calculation leads to the same result as that which has already been given above for the 50th parallel. The hemisphere which has the long winter at a time of great eccentricity has an excessive difference between summer and winter; while in the other hemisphere the difference is very slight. All this contains nothing new as regards an explanation of glacial periods, especially since this calculation refers only to one hemisphere as a whole and cannot be applied to given parallels of latitude.

Darwin's discussion of Ball's theory.—Substantially the same conclusion is reached by G. H. Darwin, who has devoted an article to a discussion of Sir Robert Ball's book.[1]

Darwin lays no special emphasis upon Ball's statement above referred to, and believes that the main point is contained in the following consideration :—

When the eccentricity is at a maximum, and winter occurs in aphelion, the length of the winter is to that of the summer as 6 : 5. The reverse is true with summer in aphelion. We therefore have the condition that the daily supply of heat during the short summer is to that during the long winter as $1/5\ (1+\alpha) : 1/6\ (1-\alpha)$. During an interglacial epoch, on the other hand, the ratio is $1/6\ (1+\alpha) : 1/5\ (1-\alpha)$. If these extreme values of the daily supply of heat are compared, it is seen that they are related to one another as

$$\frac{6(1+\alpha)}{5(1-\alpha)} : \frac{5(1+\alpha)}{6(1-\alpha)} = \frac{36}{25}.$$

The ratio of the daily supply of heat in summer to that in winter, when winter comes in aphelion, is to the same ratio during aphelion summer as 36 : 25. This is

[1] G. H. Darwin : "Review of Ball's 'The Cause of an Ice Age,'" *Nature*, XLV., 1892, 289-291 ; and "The Astronomical Theory of the Glacial Period," *ibid.*, LIII., 1895-96, 196-197.

the more emphatic expression for the relative severity of the climate at a time of maximum eccentricity, when the longest winter coincides with aphelion. It has, however, already been fully explained that this excessive annual range in solar climate does not, by any means, furnish the conditions for a glacial period. Nevertheless, the determination of the relations of these extreme conditions is of some interest in general climatology. The term, a, has a different value for each parallel of latitude, as has already been noted.

Culverwell's discussion of Ball's theory.—Culverwell[1] has discussed Ball's "astronomical theory" of glacial periods very critically, and has taken the pains, using Meech's results, to obtain the actual amounts of heat received by latitudes 40° to 80° in winter, at the height of the "Ice age," when the winter is 200 days long. If the parallels of latitude which receive the same amounts of heat from the sun during the winter are compared one with another at the height of the Ice age and at present, it appears that the latitudes which have the same "sun-heat" winter isotherms are as follows:—

LATITUDES OF SOLAR WINTER ISOTHERMS DURING AN ICE AGE AND AT PRESENT.

During an ice age, -	40°	50°	60°	70°	80° N.
At present, - -	44·2	54	63·5	74	84·5

The climatic effect of the longest winter as compared with that at present is that the latitude of 54°, for example, *now* has the same solar climate as latitude 50° would have in the supposed ice age. The change is less than that which would be experienced if London were to be moved up to the latitude of Edinburgh. Culverwell is certainly right when he says that no ice age could be produced by such a climatic displacement. "When we take account of the ocean currents it seems probable that, instead of being lowered, the winter temperature in the British Isles would be raised in the long winter of the supposed glacial epoch. For the Gulf Stream flows at about four miles (6 km.) per day between the Azores and Norway—that is about 10° of the earth's surface in six months, so that we may fairly suppose the mid-winter

[1] E. P. Culverwell: "A Mode of Calculating a Limit to the Direct Effect of Great Eccentricity of the Earth's Orbit on Terrestrial Temperatures, showing the Inadequacy of the Astronomical Theory of Ice Age and Genial Ages," *Philosoph. Mag.*, XXXVIII., 1894, 541-552; "A Criticism of the Astronomical Theory of the Ice Age and of Lord Kelvin's Suggestions in Connection with a Genial Age at the Pole," *Geol. Mag.*, N.S., XXII., 1895, 3-13, 55-65; see also "A Criticism of the Astronomical Theory of the Ice Age," by the same author, in *Nature*, LI., 1894-95, 33-35. This is a criticism of the astronomical theory of the Ice age advocated by Croll and Ball.

heating of these countries to be dependent on the summer heating at about lat. 40° – 45°. Now during the 166 days of the short summer in the epoch of great eccentricity, these latitudes received a greater daily average of heat than any latitude, even the equator, now receives in an equal time. Hence it is likely that the midwinter receipt of ocean heat in that epoch was much greater than at present."

The "astronomical theory of the Glacial period" is thus shown to be incompetent to explain the facts. It also appears, in the light of the foregoing discussion, that the known periodic changes in the elements of the earth's orbit do not warrant us in concluding that any very marked changes have taken place in terrestrial climates. In fact, from an astronomical point of view, it must rather be concluded that there has been a certain permanence of the climates of the earth's surface.

De Marchi's theory.—After a very thorough theoretical investigation of the climatic conditions during the Glacial period, and of the dependence of the temperature of the air upon the ratio of the radiant energy received to that lost by the earth, as well as upon the distribution of land and water, Luigi de Marchi comes to the conclusion that neither the astronomical nor the geological theories[1] lead to any plausible explanation of the Ice age. De Marchi does find, however, that a slight decrease from 0·6 to 0·54 in the coefficient of transmission of the atmosphere for solar rays, accompanied by a corresponding change in the coefficient of transmission for the heat radiated from land and water, would suffice to produce the climatic conditions of an ice age in the middle and the higher latitudes, and the decrease would be greater in marine than in continental climates.[2]

The difference in temperature between the continents and the oceans in higher latitudes would thus be decreased, and this, according to

[1] Elevations of the land.

[2] Luigi de Marchi finds the following changes in the mean temperatures for $q=0{\cdot}54$ and $q=0{\cdot}6$:—

Latitude, - -	10°	20°	30°	40°	50°	60°	70°	80°	90°
Marine Climate,	–0·1	–0·4	–0·9	–1·6	–2·5	–3·8	–5·0	–4·9	–4·8
Continental Climate,	–0·1	–0·5	–1·0	–1·7	–2·4	–3·0	–3·1	–2·5	–1·9

These calculated differences of temperature are probably much more likely to be right than the temperatures of the different parallels of latitude themselves, which de Marchi has obtained, and which, in the opinion of the present author, have no reality whatever. The former at least give an accurate general idea of the distribution of temperature.

Brückner, is one of the most important conditions for a rainy period on the continents, and indirectly also for a glacial advance. The temperature gradient between pole and equator is at the same time increased, and the general circulation of the atmosphere thus becomes more active. The decreased annual range of temperature in the higher latitudes would also favour the development of glaciers.

De Marchi further refers to the fact that, according to Sonklar, the glaciers of the Hohen Tauern covered 422 sq. km. at the time of their most recent maximum development; while, after 20 years of retreat, this area has, according to Brückner, been reduced to 363 sq. km. The area has therefore decreased by about one-seventh, and perhaps it may be assumed that the decrease has amounted to one-fifth. This glacial retreat corresponds to a change of climate of hardly 1° in temperature, according to Brückner. De Marchi is therefore of the opinion that a decrease of temperature of 4° – 5°, according to his calculation, together with the other favourable conditions above referred to, would probably explain the occurrence of an ice age. When, on the other hand, the coefficient of transmission again increases, there is an increase of temperature in the higher latitudes, and a more uniform distribution of temperature between the equator and pole. De Marchi does not feel prepared to say what causes may bring about an increase or a decrease in the coefficient of transmission of the earth's atmosphere, *i.e.*, an increase or a decrease in the amount of water vapour, and of carbon dioxide contained in the atmosphere.[1]

Arrhenius' theory.—Svante Arrhenius considers it much more likely that the cause of the great changes in temperature between the Miocene epoch and the Glacial period was a change in the amount of carbon dioxide in the atmosphere, and concludes that a two-fold to three-fold increase in the present amount of this gas would raise the mean temperature of the circumpolar regions 8° – 9°; while, on the other hand, a decrease of 55 to 62 per cent. of the present content could produce a fall in temperature of 4° – 5° in middle latitudes (40° – 50°), which would suffice to explain the temperature of the glacial period. Accepting the interesting conclusions reached by Högbom, Arrhenius believes that it is by no means unlikely that increased volcanic activity at different periods of the earth's history may have caused changes in the amount of carbon dioxide in the atmosphere.[2]

[1] Luigi de Marchi: *Le Cause dell' Era Glaciale*, Padua, 1895.

[2] Svante Arrhenius: "On the Influence of Carbonic Acid in the Air upon the Temperature of the Ground," *Philosoph. Mag.*, 5th Series, XLI., 237-276.

Knut Angström,[1] however, has demonstrated that the theory of Arrhenius is founded on an erroneous interpretation of observations, attributing to carbon dioxide absorption effects which are principally due to water vapour. Angström also proves that the earth's atmosphere now contains sufficient carbon dioxide to absorb substantially all of the solar radiations which are capable of being removed by this gas, rays constituting not more than $1\frac{1}{2}$ per cent. of the whole.

Chamberlin has also recently advanced a working hypothesis of a cause of glacial periods, the fundamental postulate of which is that variations in the amount of carbon dioxide in the atmosphere have been the direct cause of variations of climate. This hypothesis takes account of the changes in the amount of carbon dioxide in the atmosphere which result from the weathering of rocks and through the agency of organisms, and also considers the effects of continental elevation and denudation.[2]

Changes in the position of the earth's axis.—The simplest and most obvious explanation of great secular changes in climate, and of the former prevalence of higher temperatures in the northern circumpolar region, would be found in the assumption that the earth's axis of rotation has not always had the same position, but that it may have changed its position as the result of geological processes, such as extended rearrangements of land and water. This hypothesis has frequently been advanced, but has still more frequently, and partly for geological reasons, been shown to be untenable. The question for us to consider in this connection is simply whether mathematical physics furnishes us with any warrant for believing in the possibility of a considerable change in the position of the earth's axis.

This mathematical problem has been investigated by G. H. Darwin, who has come to the conclusion that if the earth is considered to be quite rigid, the pole may have moved about 3° from its original position. If, however, the earth is considered plastic, as it undoubtedly is to a certain degree, so that it could readjust itself to the form of equilibrium, then there is the possibility of a cumulative effect, and

[1] Knut Angström: "Ueber die Bedeutung des Wasserdampfes und der Kohlensäure bei der Absorption der Erdatmosphäre," *Annalen der Physik*, (4) III., 1900, 720-732. See also "Knut Angström on Atmospheric Absorption," *Monthly Weather Review*, XXIX., 1901, 268.

[2] T. C. Chamberlin: "Working Hypothesis of a Cause of Glacial Epochs," reviewed by Bailey Willis: "Climate and Carbonic Acid," *Pop. Sci. Mo.*, LIX., 1901, 242-256. In this connection see also N. Ekholm: "On the Variations of the Climate of the Geological and Historical Past and their Causes," *Quart. Journ. Roy. Met. Soc.*, XXVII., 1901, 1-61.

the pole may have wandered 10° – 15° from its original position. No such cumulation is possible with respect to the obliquity of the ecliptic.[1] More recently, Schiaparelli[2] has taken up the question whether, and in what way, changes in the form of the earth may affect the position of the axis of rotation. The results reached agree essentially with those of Darwin.

It cannot yet be considered as proved beyond question by astronomical and mechanical considerations that the geographical pole has remained in the same position on the earth's surface. The permanence of the pole may be a fact to-day, but it must also be an established fact so far as the earth's past history is concerned. A permanent position of the earth's pole is possible only in an earth of a certain rigidity. Geological processes, although they be insignificant, may, if they only operate long enough, interfere with the conditions of the permanence of the pole, even after these conditions have once been attained. Thus considerable movements of the pole may be brought about as long as the earth is not absolutely rigid.

The climatic consequences which would result from a displacement of the north pole to latitude 70° N., and longitude 20° W., have recently been discussed by Davis, in a highly instructive paper.[3] The change in the limits of the wind and rain belts, which would follow such a displacement of the pole, would tend to glaciate northwestern Europe and northwestern America; would place arid trade wind climates on the northern side of the belt now occupied by the equatorial rains of Africa and of South America, and would at the same time place the equatorial rains on the northern margin of the arid land areas now found in the southern parts of these continents. It is suggested that evidence of such changes in climate should be sought in the topographic forms and in the conditions of lake desiccation, or overflow, in torrid Africa and South America.

[1] G. H. Darwin: "On the Influence of Geological Changes on the Earth's Axis of Rotation," *Proc. Roy. Soc.*, XXV., 1876-77, 328-332; also *Nature*, XV., 1876-77, 360-361. S. Haughton: "Preliminary Formulae relating to the Internal Change of Position of the Earth's Axis arising from Elevations and Depressions caused by Geological Changes," *Proc. Roy. Soc.*, XXVI., 1877, 51-55; also *Nature*, XV., 1876-77, 542-543.

[2] G. V. Schiaparelli: "De la Rotation de la Terre sous l'Influence des Actions Geólogiques," *Mémoire présenté à l'Observatoire de Poulkova*, St. Petersburg, 1889. Reviewed in *Pet. Mitt.*, XXVIII., 1892, 42-45.

[3] W. M. Davis: "A Speculation in Topographical Climatology," *Am. Met. Jour.*, XII., 1895-96, 372-381.

Changes in the distribution of land and water.—Some idea of the great climatic changes, which would result from a radically different distribution of land and water on the earth's surface, may be gained from what has already been said concerning marine and continental climates. Lyell was the latest advocate of great geographic rearrangements as a possible cause of climatic change.[1] Von Kerner has also made a speculation of a similar kind (see p. 208).

Secular variations of climate.—The question whether the climate of certain districts has, or has not, changed within historical times, has been the subject of numerous investigations and discussions, but no definite results have been reached. It has often been thought that a change in the climate of a certain country could be proved beyond question; but, on the other hand, it has just as often been found possible to prove that these changes have not taken place. Thus, for example, it has frequently been plausibly maintained that there was once a more abundant water supply in northern Africa. This has been disproved by Partsch, who has shown that such could hardly have been the case within historical times. Partsch bases his argument on the fact that the ancient settlements on the interior lakes of northern Africa, which do not overflow and which may be regarded as real rain-gauges, clearly show that these lakes formerly contained no more water than they do at present. An examination of the records of temperature and of rainfall from the period at which they first began to be kept, and for which they are to some extent comparable—*i.e.*, at the most, during 150 years—does not bring to light any evidence of a progressive change in the temperature of the air, or in the amount of rainfall. Whenever it has been supposed that these records did show an increase in temperature or in rainfall, it has always turned out that this increase may have been due to the method of exposing the thermometer, or the rain-gauge. The oldest meteorological records were always begun in cities, which grew very rapidly in course of time, and it has already been pointed out that "city temperatures" are always higher than those in the surrounding country. Furthermore, little attention was formerly paid to the influence of the height of the rain-gauge upon the amount of rain collected. Most of the older measurements of rainfall, which were obtained in cities by exposing the gauges on terraces and roofs, or within enclosed court-yards, gave too small a rainfall. To this extent, therefore, do direct meteorological records leave us in doubt as to the question of a change in the climatic elements.

[1] Sir C. Lyell: *Principles of Geology*, Vol. I., Chap. XII.

A great many apparent indications of a progressive change of climate, either as regards temperature or rainfall, are undoubtedly to be referred to periodic fluctuations in the mean values of the climatic elements, from which too general conclusions have been drawn, or which have been interpreted from too one-sided a point of view. Dufour, who has made a thorough study of the available evidence concerning a change in climate, concludes that the uncertainties connected with this evidence make it impossible to regard a change of climate as proved. The question, nevertheless, remains an open one, and the common assertion that the climate is not changing is, under the circumstances, a no more legitimate consequence of known facts than is the opposite view.[1]

At the conclusion of a careful compilation and discussion of the dates of the vintage in France, Angot says that these data, which extend back into the 14th century, give no support to the view, held by so many persons, that there is a progressive change of the climate for the worse. The dates of the vintages seem rather to point toward oscillations of the climatic elements. During the period 1775-1875, the average date of the grape harvest at Aubonne was 10 days earlier than the average date in the preceding century, and only three days later than that recorded two centuries before. At the present time, the date of the grape harvest in Aubonne is again exactly the same as at the close of the 16th century.[2] The average date of the vintage at Dijon shows the following secular variations: 14th century (mean of 13 years only), October 25; 15th century (6 decades), October 25; 16th century, October 28; 17th century, October 24·5; 18th century, October 28·8; 19th century (8 decades), October 30.

In the United States, all the older temperature records and rainfall records have been collected, reduced, and carefully discussed by Schott. In his chapter on the "Secular Variation of Air Temperature," the author concludes, on the basis of the longest series of observations taken from Maine to California, that these records show variations of temperature, occurring over a great extent of territory with the same characteristics and with considerable uniformity. These variations have the characteristics of irregular waves, which represent a succession of warmer and of colder periods, but during which the temperature varies only 0·5°-1·0° to one side or the other of the mean.

[1] L. Dufour: "Notes sur le Problème de la Variation du Climat," *Bull. Soc. Vaud.*, X., Lausanne, 1870.

[2] A. Angot: "Étude sur les Vendanges en France," *Ann. Bur. Centr. Met.*, 1883, I. (Paris, 1885), B. 29-120

There seems, however, to be nothing in these variations which could lead to the view that there is a progressive change in any one direction. The maxima and the minima of temperature follow one another at intervals of about 22 years on the Atlantic coast; while the temperature waves of the interior States seem to be shorter, the interval between the maxima and the minima being about seven years. These undulations are, however, much too irregular and too uncertain to serve as a basis for forecasting.[1]

The United States seem to offer the most favourable conditions for answering the question as to the extent to which the increasing cultivation of large districts of country may result in a change of climate. In the east, there has been an extraordinary decrease in the amount of territory formerly covered by forests; while, on the other hand, a good deal of planting has been done on the western prairies and plateaus. No corresponding change in temperature, or in precipitation, has, however, thus far been demonstrable.[2] Whitney does not believe that any improvement in the climate of the arid west is to be expected as the result of man's agency.[3]

A general review of the present status of the question as to changes of climate is given by Brückner, in the first chapter of his great work, *Klimaschwankungen* (Vienna, 1890). The same author has given a bibliography of the literature on changes of climate in *Geographisches Jahrbuch*, Vol. XV., 1891, pp. 438-440, and Vol. XVII., 1894, 347-350.

[1] C. A. Schott: "Tables, Distribution and Variations of the Atmospheric Temperature in the United States and some Adjacent Parts of America," *Smithsonian Contr. to Knowledge*, XXI., 1876, 302-320; "Tables and Results of the Precipitation, in Rain and Snow, in the United States," 2nd edition, *ibid.*, XXIV., 1885, Art. II., V.-XX., 1-249.

[2] See M. W. Harrington: "Rainfall and Snow of the United States," *U.S. Weather Bureau, Bull. C*, 1894, Text, 19-20; Atlas, Sheet XXII., Map 4.

[3] J. D. Whitney: "Brief Discussion of the Question whether Changes of Climate can be brought about by the Agency of Man, and on Secular Climatic Changes in General, with Special Reference to the Arid Region of the United States," *The United States*, Supplement I., Boston, 1894, Appendix B, 290-317.

CHAPTER XXII.

PERIODIC VARIATIONS OF CLIMATE.

Oscillations of climate.—The discovery that the climatic elements are subject to certain oscillations, first remaining for a time above the mean derived from many years of observations, and then again falling below this mean, has already led to many investigations which have been undertaken with a view to determining whether or not these oscillations occur with some regularity, in definite cycles. The problem therefore involves the determination of the length of the period and the amplitude of these oscillations. Such an investigation may be taken up from two different points of view. On the one hand, we may assume a period of a given length, which is determined by certain physical considerations, and we may then try to discover whether the successive occurrences of the various phenomena in question actually show the existence of such a period. On the other hand, we may start without any definite period in mind, and may try to discover the unknown length of the period by different combinations of the successive values of one meteorological element. The method first mentioned, is the one usually employed, because it naturally forces itself upon the inquiring mind. Thus, from the very earliest times, persons have tried to prove that the moon's phases, and especially the varying positions of the moon with reference to the earth, correspond to certain changes in the meteorological elements, but, as is well known, none of these attempts has yet led to any noteworthy result.

Sunspots and meteorological cycles.—After the periodicity in the occurrence of sunspots had been recognised, it naturally seemed to be a very fruitful problem to prove that there are also corresponding periods in the succession of the meteorological elements. The apparent course of the sun controls all accepted meteorological cycles, and it is

therefore very natural to ascribe to the visible changes on the surface of the sun a noticeable influence upon our atmosphere. Nevertheless, the results of the numerous and varied studies which have been made of this problem have not quite come up to expectations. The influence of sunspots upon the meteorological elements has turned out to be rather insignificant. Even in the most marked cases, the only thing that can be considered as proved is that there are traces of a parallelism in the march of certain meteorological elements and that of the sunspot period. A forecast of the weather on the basis of sunspot periodicity is still impracticable. The case is, however, quite different with the magnetic elements and their variations, for these are subject to the same cycles as those of the sunspots, with great ranges.

A full discussion of the results of the investigations which have been made of the 10-11 year periodicity of the climatic elements, following the sunspot period, cannot be entered into here.[1] We must content ourselves with citing a few cases only.

Sunspots and temperature.—Köppen has given the clearest proof of the existence of a sunspot period in the mean annual temperatures of the different climatic districts on the earth's surface.[2] In the tropics, the parallelism between the variations of the mean annual temperature and those of the sunspot frequency is fairly well marked, but it is less apparent in middle and in higher latitudes. The mean value of the variation in the mean annual temperatures, from a sunspot minimum to a sunspot maximum, amounts to 0·73° in the tropics, and to 0·54° in the extra-tropical zones. The following data, which show the departures of the annual mean temperatures from the means derived from many years' observations, will serve to show the march of this phenomenon in the tropics:—

SUNSPOT PERIODS IN THE MEAN ANNUAL TEMPERATURES OF THE TROPICS.

Sunspot Minimum,	- -	1	2	3	4 years.
+0·33°		+0·15°	+0·04°	−0·21°	−0·28°
Sunspot Maximum,	1	2	3	4	5 years.
−0·32°	−0·27°	−0·14°	+0·08°	+0·30°	+0·41°

[1] The earlier results may be found clearly summarised in H. Fritz: *Die Beziehungen der Sonnenflecken zu den magnetischen und meteorologischen Erscheinungen der Erde*, Harlem, 1878, 275 quarto pages, with tables.

[2] W. Köppen: "Ueber mehrjährige Perioden der Witterung," *Z.f.M.*, VIII., 1873, 241-248, 257-267; XV., 1880, 279-283; XVI., 1881, 140-150.

The maximum temperature occurs about 0·9 year before the sunspot minimum, and the minimum temperature occurs almost simultaneously with the sunspot maximum.

Sunspots and rainfall.—Meldrum and Lockyer, in particular, have endeavoured to prove that the sunspot period has an influence upon the annual rainfall, the results of their investigations going to show that more rain seems to fall at a period of maximum sunspot activity than at a sunspot minimum. A regular variation in the rainfall appears more distinctly in the tropics than in the higher latitudes. In general, however, the excess of rainfall at a time of sunspot maxima appears only in the larger figures, *i.e.*, in the means, and in the case of the majority of stations. It by no means appears in every period and at every individual station. Indeed, some rainfall records, which extend over many years, actually give contradictory results.[1] For this reason, a practical application of the result is out of the question. There has been a hope that some use could be made of the recognised sunspot cycle in the amount of rainfall in India. In fact, it has been thought that the years of famine, which are connected with a failure of the monsoon rains, or with a deficiency in them, can be connected with the sunspot cycle.[2]

Recently the Lockyers have made an important study of the variations in rainfall in the region surrounding the Indian Ocean in the light of solar changes in temperature.[3] The following are among the conclusions reached:—"It has been found from a discussion of the chemical origin of lines most widened in sunspots at maxima and minima periods, that there is a considerable rise above the mean temperature of the sun around the years of sunspot maximum and a considerable fall around the years of sunspot minimum. It has been found from the actual facts of rainfall in India (during the southwest monsoon) and Mauritius, between the years 1877 and 1886, as given by Blanford and Meldrum, that the effects of these solar changes are felt in India at sunspot maximum, and in Mauritius at sunspot minimum.

[1] F. W. Very has pointed out that opposite effects are to be anticipated in different terrestrial regions from increase of solar radiation ("The Variation of Solar Radiation," *Astrophysical Journal*, VII., 1898, 255-272).

[2] J. N. Lockyer and W. W. Hunter: "Sunspots and Famines," *Nineteenth Century*, II., 1877, 583-602.

[3] Sir Norman and W. J. S. Lockyer: "On Solar Changes of Temperature and Variations in Rainfall in the Region surrounding the Indian Ocean." Read before the Royal Society, November 22, 1900, *Nature*, LXIII., 1900, 107-109, 128-133.

Of these the greater is that produced in the Mauritius at sunspot minimum. The pulse at Mauritius at sunspot minimum is also felt in India, and gives rise generally to a secondary maximum in India. India therefore has two pulses of rainfall, one near the maximum and the other near the minimum of the sunspot period. It has been found from a study of the Famine Commission reports that all the famines therein recorded which have devastated India during the last half-century (the investigation has not been carried further back) have occurred in the intervals between the two pulses." The authors believe that if as much had been known in 1836 as we know now, the probability of famines at all the subsequent dates might have been foreseen.

Archibald believes that he has found in the winter rains of northern India a periodicity the opposite of that of the sunspot cycle, *i.e.*, more rain at the time of sunspot minima. Independently of Archibald, Hill tried to show that in the years of sunspot maxima the summer monsoon rainfall is above the mean, but that the winter rainfall of northern India is then deficient. The reverse is said to be true at a time of sunspot minima.[1] But even Blanford himself, who was so much interested in the meteorological sunspot cycle, and who maintained that it existed, came to the following conclusion in 1889 :—[2]

"Much has been written of late years of the supposed variation of rainfall in a cycle of 11 years, coincidently with the now well-known variation of sunspots. Without venturing to pronounce any opinion on the question whether any such variation exists in the rainfall of the globe generally, it is certain that that of India, as a whole, has not displayed it during the last 22 years. But while this is true of the country *as a whole*, the rainfall of the Carnatic, which chiefly occurs at a later season than the heavy rains of other parts of India, certainly showed a somewhat striking fluctuation in the 11 years, 1864-1874, and again, though less regular, in the next 11 years, 1875-1885. It is still a matter of opinion whether this was fortuitous or otherwise, a question which can be decided only by further experience."

Sunspots and tropical cyclones.—At the meeting of the British Association for the Advancement of Science, held at Brighton in 1872, Meldrum first called attention to the fact that the cyclones of the Indian Ocean between the equator and latitude 25° S., seem to be less

[1] S. A. Hill: "Variations of Rainfall in Northern India," *Ind. Met. Mem.*, I., No. VII.

[2] H. F. Blanford: *The Climates and Weather of India, Ceylon, and Burmah*, London and New York, 1889, p. 80.

frequent in the years of sunspot minima than in the years of sunspot maxima. The violence of these cyclones is also greater at the time of sunspot maxima. Pöey later made a similar study of the cyclones of the Antilles, with reference to their periodicity, and came to the same conclusion as that reached by Meldrum.[1] In both the Atlantic and the Indian Ocean, cyclones occur more frequently with increasing numbers of sunspots, and become less frequent again at times of sunspot minima.

Sunspots and other meteorological phenomena.—Furthermore, the minimum winter temperatures, as well as severe winters themselves; the periods of the advance and retreat of glaciers; the height of rivers; in short, all the meteorological elements and the phenomena which depend upon them, have been studied in their relation to sunspots, and evidences of some connection have been found. This evidence is, however, still too uncertain to show any actual relation of cause and effect.[2]

Brückner's oscillations of climate.[3]—By following the second of the two methods above referred to, Brückner came to the conclusion that there is a 35-year period of climatic oscillation. A study of the remarkable long-period oscillations in the water level of the Caspian Sea, forced Brückner to the conclusion that these changes have a period of 34 to 36 years. Out of this discovery there naturally grew an investigation into a possible correspondence of periodicities in temperature and precipitation and in the levels of the rivers which flow into the Caspian Sea, as well as in the dates of the opening and closing of the rivers of the Russian empire given in Rykatschew's great work.[4] This study shows that the cause of the periodic oscillations in the level of the Caspian Sea is to be found in corresponding

[1] C. Meldrum: "On a Periodicity in the Frequency of Cyclones in the Indian Ocean south of the Equator," *Rept. Brit. Ass.*, 1872, 56-58; "On a Periodicity of Cyclones and Rainfall in Connection with the Sunspot Periodicity," *ibid.*, 1873, 466-478. A. Pöey: "Rapports entre les Taches solaires, les Orages à Paris et à Fécamp, les Tempêtes et les Coups de Vents dans l'Atlantique nord," *Comptes Rendus*, LXXVII., 1873, 1223-1226.

[2] See also F. G. Hahn: *Ueber die Beziehungen der Sonnenfleckenperiode zu den meteorologischen Erscheinungen*, Leipzig, 1877. D. Ragona: "Ueber die Jährliche Periode der Variabilität der Temperatur," *Z.f.M.*, XIII., 1878, 33-35. H. C. Russell believes that there is a 19-year periodicity in the weather conditions of New South Wales ("Periodicity of Good and Bad Seasons," *Proc. Roy. Soc. N.S. Wales*, 1896; abridged in *Nature*, LIV., 1896, 379-380).

[3] E. Brückner: *Klimaschwankungen seit* 1700, *nebst Bemerkungen über die Klimaschwankungen der Diluvialzeit*, Vienna, 1890.

[4] M. Rykatschew: *Rep.f.Met.*, II. Supplementary Volume, St. Petersburg, 1887.

periods in the temperature and the rainfall. This led to the conclusion that all European Russia has passed through great climatic oscillations since the beginning of the eighteenth century. Cold, wet periods prevailed about the years 1745, 1775, 1810, 1845, and 1880; and dry, warm periods prevailed about the years 1715, 1760, 1795, 1825, and 1860. These climatic oscillations affected the rivers by determining the length of time during which they were frozen, and by controlling the depth of water in them; and the rivers, in their turn, affected the great Caspian Sea by first raising and then lowering its level.

Further investigations into the oscillations of lakes without outlets in general, as well as of rivers and discharging lakes, and then into the oscillations of the rainfall of the whole world, led Brückner to the conclusion that the period discovered in the case of the Caspian Sea holds over a much wider area. The periods of highest and lowest levels in lakes were as follows:—Minima in 1720, 1760, 1798, 1835, 1865; maxima in 1740, 1777, 1820, 1850, 1880. The rainfall periods after 1830 have been as follows:—Dry periods, common to all parts of the world, in 1831-1840, and 1861-1865; wet periods in 1846-1855, and in 1876-1880. Districts certainly exist in which these phases and epochs are exactly reversed. These are called "districts which are permanently exceptional" by Brückner, and they are limited almost altogether to the regions which have marine climates.

Brückner[1] has lately made a supplementary study of the latest rainfall data for Russia and the United States, as well as for some stations in Central Europe and Eastern Siberia, and finds satisfactory confirmation of his earlier conclusions, in a decrease in the rainfall for these districts as a whole, beginning about the middle of the decade 1880-1890.

Hann[2] has recently discussed the monthly and yearly means of the rainfall at Padua (1725-1900), Klagenfurt (1813-1900), and Milan (1764-1900), and finds that they all conform to a period of definite length, and that the alternation of wet and dry epochs is in harmony with the 35-year period of rainfall indicated by Brückner.

Oscillations in the mean annual pressure.—The study of the oscillations in the mean annual pressure led to the important discovery that the periods of these oscillations are not the same on the continents as over the oceans, but that there is a compensating relation between

[1] E. Brückner: "Zur Frage der 35-jährigen Klimaschwankungen," *Pet. Mitt.*, XLVIII., 1902, 173-178.

[2] J. Hann: "Ueber die Schwankungen der Niederschlagsmengen in grösseren Zeiträumen," *S. W.A.*, CXI., 2a, 1902, 1-120.

them. This throws some light upon the origin of the districts above-mentioned as being exceptional in their rainfall conditions. The march of the pressure over the oceans and over the continents reflects the oscillations in the rainfall. High pressure over the ocean involves a dry period there, while at the same time the low pressure over the land causes a rainy period there, and *vice versa*.

Another important result which was reached by Brückner is that every rainy period is accompanied by a decrease in all differences of pressure, and that every dry period is accompanied by an increase in these differences. This statement is true not only in point of space, *i.e.*, of the gradient, but also in point of time, in relation to the annual variation of pressure. The dry periods on the Eurasian continent are characterised by a further deepening of the barometric depression over the North Atlantic Ocean; an increase of the ridge of high pressure which extends from the Azores northeast across central Europe to Russia; a deepening of the trough of low pressure over the northern portion of the Indian Ocean and the southern portion of the China Sea; a weakening of the mean annual pressure in the anticyclone over Siberia; and by a general increase in the amplitude of the annual range.

Departures of temperature and of rainfall during the different cycles.—The fluctuations in the rainfall are much more marked in the interior of the continents than on the coasts. In western Siberia, more than twice as much rain may fall in the wet, as in the dry periods. The general average amplitude of the fluctuation is only 12 per cent., but excluding the exceptional districts, it is 24 per cent. The mean temperatures also are subject to the same periodicity. The mean of 280 stations, scattered all over the world, shows that the years, 1791-1805, 1821-1835, 1851-1870, were periods of high temperature, and that the periods of low temperature were 1806-1820, 1836-1850, and 1871-1880. The amplitude of the temperature ranges for the whole world is, in round numbers, 1°, which is therefore greater than that which corresponds to the sunspot period. The mean temperature departures for the whole world are as follows:—

TEMPERATURE DEPARTURES DURING PERIODS OF HIGH AND LOW TEMPERATURES.

1736-1740, - - -	−0·43	1821-1825, - - -	+0·56
1746-1750, - - -	+0·45	1836-1840, - - -	−0·39
1766-1770, - - -	−0·42	1851-1855, - - -	+0·11
1791-1795, - - -	+0·46	1866-1870, - - -	+0·11
1811-1815, - - -	−0·46	1881-1885, - - -	−0·08

These departures relate to the stations for which records are now available, and naturally do not really represent the temperature conditions of the whole world, especially not during the earlier periods. The average length of the periods is seen to be 36 years. A period of about 35 years, which is now regarded by Brückner as being more nearly the true length of the period, is likewise apparent in the records which reach back much further than do those of temperature. These records are those of the conditions of ice in rivers (from the year 1736); the date of vintage (from the year 1400), and the occurrence of severe winters (from the year 800). Lockyer believes that this thirty-five year period in climate is connected with a long-period sunspot variation of thirty-five years.[1]

Variations in the Swiss glaciers.—The periodicity in climate discovered by Brückner is strongly supported by the results of Richter's study of the variations in the Alpine glaciers.[2] This study shows that "the advance of the glaciers recurs in periods whose length varies between 20 and 45 years, and was exactly 35 years in the mean of the last three centuries. These periods, on the whole, agree with the dates which Brückner obtained for his climatic oscillations during the last three centuries. The advance of the glaciers is noticeable during the cold, damp period." The years when the glaciers began to advance were 1592, 1630, 1675, 1712, 1735, 1767, 1814, 1835 and 1875; and the cold periods, according to Brückner, were 1591-1600, 1611-1635, 1646-1665, 1691-1715, 1730-1750, 1766-1775, 1806-1820 and 1836-1855.

Price of grain and climatic cycles.—The effect which these oscillations of climate have upon the harvest and upon the price of grain in Europe has recently been fully discussed by Brückner in a special paper.[3] He comes to the conclusion that the prices of grain were on the average 13 per cent. higher in the wettest lustrum than in the driest.

General conclusion regarding supposed changes of climate.—Brückner's oscillations of climate help to explain the prevailing views, which are so often contradictory, of a change in the climate of certain districts for better or worse to a moister, or a drier, condition. Such views have

[1] W. J. S. Lockyer: "The Solar Activity, 1833-1900," Proc. Royal Soc., LXVIII., 1901, 285. Abstract in *Nature*, LXIV., 1901, 196-197.

[2] E. Richter: "Geschichte der Schwankungen der Alpengletscher," *Zeitschr. deutsch. österr. Alpenver.* (Vienna), XXII., 1891, 1-74.

[3] E. Brückner: "Der Einfluss der Klimaschwankungen auf die Ernteerträge und Getreidepreise in Europa," *Geograph. Zeitschr.*, I., 1895, 39-51, 100-108.

grown up as the result of the impression made by different phases of these oscillations. The improvement in the climate in the western portion of the United States has, according to Brückner, been associated with a wet period of his climatic oscillations. A drier phase, which began about 1886, ended this, as has been the case in Egypt and Siberia, where not only the returns of agriculture, but the extent of the land which is available for farming, fluctuate directly in sympathy with the oscillations of climate. Continental areas are just the ones which are most markedly affected by these changes.[1]

Brückner expresses no opinion concerning the probable causes which operate to bring about this 35-year periodicity in climatic oscillations. He is properly content with having demonstrated the existence of the period with a high degree of probability.

[1] In obtaining the means of the different climatic elements it would be well to see to it that these means are based upon long series of years which embrace a whole period of climatic oscillations, or several such periods.

APPENDICES.

A. CONVERSION TABLES.

B. ADDENDA.

TABLE I.

CENTIGRADE SCALE TO FAHRENHEIT.

Centigrade.	·0	·1	·2	·3	·4	·5	·6	·7	·8	·9
°	°F.	°F.	°F.	°F.	°F.	°F	°F.	°F.	°F.	°F.
+50	+122·00	+122·18	+122·36	+122·54	+122·72	+122·90	+123·08	+123·26	+123·44	+123·62
49	120·20	120·38	120·56	120·74	120·92	121·10	121·28	121·46	121·64	121·82
48	118·40	118·58	118·76	118·94	119·12	119·30	119·48	119·66	119·84	120·02
47	116·60	116·78	116·96	117·14	117·32	117·50	117·68	117·86	118·04	118·22
46	114·80	114·98	115·16	115·34	115·52	115·70	115·88	116·06	116·24	116·42
+45	+113·00	+113·18	+113·36	+113·54	+113·72	+113·90	+114·08	+114·26	+114·44	+114·62
44	111·20	111·38	111·56	111·74	111·92	112·10	112·28	112·46	112·64	112·82
43	109·40	109·58	109·76	109·94	110·12	110·30	110·48	110·66	110·84	111·02
42	107·60	107·78	107·96	108·14	108·32	108·50	108·68	108·86	109·04	109·22
41	105·80	105·98	106·16	106·34	106·52	106·70	106·88	107·06	107·24	107·42
+40	+104·00	+104·18	+104·36	+104·54	+104·72	+104·90	+105·08	+105·26	+105·44	+105·62
39	102·20	102·38	102·56	102·74	102·92	103·10	103·28	103·46	103·64	103·82
38	100·40	100·58	100·76	100·94	101·12	101·30	101·48	101·66	101·84	102·02
37	98·60	98·78	98·96	99·14	99·32	99·50	99·68	99·86	100·04	100·22
36	96·80	96·98	97·16	97·34	97·52	97·70	97·88	98·06	98·24	98·42
+35	+95·00	+95·18	+95·36	+95·54	+95·72	+95·90	+96·08	+96·26	+96·44	+96·62
34	93·20	93·38	93·56	93·74	93·92	94·10	94·28	94·46	94·64	94·82
33	91·40	91·58	91·76	91·94	92·12	92·30	92·48	92·66	92·84	93·02
32	89·60	89·78	89·96	90·14	90·32	90·50	90·68	90·86	91·04	91·22
31	87·80	87·98	88·16	88·34	88·52	88·70	88·88	89·06	89·24	89·42
+30	+86·00	+86·18	+86·36	+86·54	+86·72	+86·90	+87·08	+87·26	+87·44	+87·62
29	84·20	84·38	84·56	84·74	84·92	85·10	85·28	85·46	85·64	85·82
28	82·40	82·58	82·76	82·94	83·12	83·30	83·48	83·66	83·84	84·02
27	80·60	80·78	80·96	81·14	81·32	81·50	81·68	81·86	82·04	82·22
26	78·80	78·98	79·16	79·34	79·52	79·70	79·88	80·06	80·24	80·42
+25	+77·00	+77·18	+77·36	+77·54	+77·72	+77·90	+78·08	+78·26	+78·44	+78·62
24	75·20	75·38	75·56	75·74	75·92	76·10	76·28	76·46	76·64	76·82
23	73·40	73·58	73·76	73·94	74·12	74·30	74·48	74·66	74·84	75·02
22	71·60	71·78	71·96	72·14	72·32	72·50	72·68	72·86	73·04	73·22
21	69·80	69·98	70·16	70·34	70·52	70·70	70·88	71·06	71·24	71·42
+20	+68·00	+68·18	+68·36	+68·54	+68·72	+68·90	+69·08	+69·26	+69·44	+69·62
19	66·20	66·38	66·56	66·74	66·92	67·10	67·28	67·46	67·64	67·82
18	64·40	64·58	64·76	64·94	65·12	65·30	65·48	65·66	65·84	66·02
17	62·60	62·78	62·96	63·14	63·32	63·50	63·68	63·86	64·04	64·22
16	60·80	60·98	61·16	61·34	61·52	61·70	61·88	62·06	62·24	62·42
+15	+59·00	+59·18	+59·36	+59·54	+59·72	+59·90	+60·08	+60·26	+60·44	+60·62
14	57·20	57·38	57·56	57·74	57·92	58·10	58·28	58·46	58·64	58·82
13	55·40	55·58	55·76	55·94	56·12	56·30	56·48	56·66	56·84	57·02
12	53·60	53·78	53·96	54·14	54·32	54·50	54·68	54·86	55·04	55·22
11	51·80	51·98	52·16	52·34	52·52	52·70	52·88	53·06	53·24	53·42
+10	+50·00	+50·18	+50·36	+50·54	+50·72	+50·90	+51·08	+51·26	+51·44	+51·62
9	48·20	48·38	48·56	48·74	48·92	49·10	49·28	49·46	49·64	49·82
8	46·40	46·58	46·76	46·94	47·12	47·30	47·48	47·66	47·84	48·02
7	44·60	44·78	44·96	45·14	45·32	45·50	45·68	45·86	46·04	46·22
6	42·80	42·98	43·16	43·34	43·52	43·70	43·88	44·06	44·24	44·42
+ 5	+41·00	+41·18	+41·36	+41·54	+41·72	+41·90	+42·08	+42·26	+42·44	+42·62
4	39·20	39·38	39·56	39·74	39·92	40·10	40·28	40·46	40·64	40·82
3	37·40	37·58	37·76	37·94	38·12	38·30	38·48	38·66	38·84	39·02
2	35·60	35·78	35·96	36·14	36·32	36·50	36·68	36·86	37·04	37·22
1	33·80	33·98	34·16	34·34	34·52	34·70	34·88	35·06	35·24	35·42
+ 0	+32·00	+32·18	+32·36	+32·54	+32·72	+32·90	+33·08	+33·26	+33·44	+33·62

TABLE I.

CENTIGRADE SCALE TO FAHRENHEIT.

Centigrade.	·0	·1	·2	·3	·4	·5	·6	·7	·8	·9
°	°F.	°F.	°F.	°F.	°F.	°F.	°F.	°F.	°F.	°F.
− 0	+ 32·00	+ 31·82	+ 31·64	+ 31·46	+ 31·28	+ 31·10	+ 30·92	+ 30·74	+ 30·56	+ 30·38
1	30·20	30·02	29·84	29·66	29·48	29·30	29·12	28·94	28·76	28·58
2	28·40	28·22	28·04	27·86	27·68	27·50	27·32	27·14	26·96	26·78
3	26·60	26·42	26·24	26·06	25·88	25·70	25·52	25·34	25·16	24·98
4	24·80	24·62	24·44	24·26	24·08	23·90	23·72	23·54	23·36	23·18
− 5	+ 23·00	+ 22·82	+ 22·64	+ 22·46	+ 22·28	+ 22·10	+ 21·92	+ 21·74	+ 21·56	+ 21·38
6	21·20	21·02	20·84	20·66	20·48	20·30	20·12	19·94	19·76	19·58
7	19·40	19·22	19·04	18·86	18·68	18·50	18·32	18·14	17·96	17·78
8	17·60	17·42	17·24	17·06	16·88	16·70	16·52	16·34	16·16	15·98
9	15·80	15·62	15·44	15·26	15·08	14·90	14·72	14·54	14·36	14·18
− 10	+ 14·00	+ 13·82	+ 13·64	+ 13·46	+ 13·28	+ 13·10	+ 12·92	+ 12·74	+ 12·56	+ 12·38
11	12·20	12·02	11·84	11·66	11·48	11·30	11·12	10·94	10·76	10·58
12	10·40	10·22	10·04	9·86	9·68	9·50	9·32	9·14	8·96	8 78
13	8·60	8·42	8·24	8·06	7·88	7·70	7·52	7·34	7·16	6·98
14	6·80	6·62	6·44	6·26	6·08	5·90	5·72	5·54	5·36	5·18
− 15	+ 5·00	+ 4·82	+ 4·64	+ 4·46	+ 4·28	+ 4·10	+ 3·92	+ 3·74	+ 3·56	+ 3·38
16	+ 3·20	+ 3·02	+ 2·84	+ 2·66	+ 2·48	+ 2·30	+ 2·12	+ 1·94	+ 1·76	+ 1·58
17	+ 1·40	+ 1·22	+ 1·04	+ 0·86	+ 0·68	+ 0·50	+ 0·32	+ 0·14	− 0·04	− 0·22
18	− 0·40	− 0·58	− 0·76	− 0·94	− 1·12	− 1·30	− 1·48	− 1·66	− 1·84	− 2·02
19	− 2·20	− 2·38	− 2·56	− 2·74	− 2·92	− 3·10	− 3·28	− 3·46	− 3·64	− 3·82
− 20	− 4·00	− 4·18	− 4·36	− 4·54	− 4·72	− 4·90	− 5·08	− 5·26	− 5·44	− 5·62
21	5·80	5·98	6·16	6·34	6·52	6·70	6·88	7·06	7·24	7·42
22	7·60	7·78	7·96	8·14	8·32	8·50	8·68	8·86	9·04	9·22
23	9·40	9·58	9·76	9·94	10·12	10·30	10·48	10·66	10·84	11·02
24	11·20	11·38	11·56	11·74	11·92	12·10	12·28	12·46	12·64	12·82
− 25	− 13·00	− 13·18	− 13·36	− 13·54	− 13·72	− 13·90	− 14·08	− 14·26	− 14·44	− 14·62
26	14·80	14·98	15·16	15·34	15·52	15·70	15·88	16·06	16·24	16·42
27	16·60	16·78	16·96	17·14	17·32	17·50	17·68	17·86	18·04	18·22
28	18·40	18·58	18·76	18·94	19·12	19·30	19·48	19·66	19·84	20·02
29	20·20	20·38	20·56	20·74	20·92	21·10	21·28	21·46	21·64	21·82
− 30	− 22·00	− 22·18	− 22·36	− 22·54	− 22·72	− 22·90	− 23·08	− 23·26	− 23·44	− 23·62
31	23·80	23·98	24·16	24·34	24·52	24·70	24·88	25·06	25·24	25·42
32	25·60	25·78	25·96	26·14	26·32	26·50	26·68	26·86	27·04	27·22
33	27·40	27·58	27·76	27·94	28·12	28·30	28·48	28·66	28·84	29·02
34	29·20	29·38	29·56	29·74	29·92	30·10	30·28	30·46	30·64	30·82
− 35	− 31·00	− 31·18	− 31·36	− 31·54	− 31·72	− 31·90	− 32·08	− 32·26	− 32·44	− 32·62
36	32·80	32·98	33·16	33·34	33·52	33·70	33·88	34·06	34·24	34·42
37	34·60	34·78	34·96	35·14	35·32	35·50	35·68	35·86	36·04	36·22
38	36·40	36·58	36·76	36·94	37·12	37·30	37·48	37·66	37·84	38·02
39	38·20	38·38	38·56	38·74	38·92	39·10	39·28	39·46	39·64	39·82
− 40	− 40·00	− 40·18	− 40·36	− 40·54	− 40·72	− 40·90	− 41·08	− 41·26	− 41·44	− 41·62
41	41·80	41·98	42·16	42·34	42·52	42·70	42·88	43·06	43·24	43·42
42	43·60	43·78	43·96	44·14	44·32	44·50	44·68	44·86	45·04	45·22
43	45·40	45·58	45·76	45·94	46·12	46·30	46·48	46·66	46·84	47·02
44	47·20	47·38	47·56	47·74	47·92	48·10	48·28	48·46	48·64	48·82
− 45	− 49·00	− 49·18	− 49·36	− 49·54	− 49·72	− 49·90	− 50·08	− 50·26	− 50·44	− 50·62
46	50·80	50·98	51·16	51·34	51·52	51·70	51·88	52·06	52·24	52·42
47	52·60	52·78	52·96	53·14	53·32	53·50	53·68	53·86	54·04	54·22
48	54·40	54·58	54·76	54·94	55·12	55·30	55·48	55·66	55·84	56·02
49	56·20	56·38	56·56	56·74	56·92	57·10	57·28	57·46	57·64	57·82
− 50	− 58·00	− 58·18	− 58·36	− 58·54	− 58·72	− 58·90	− 59·08	− 59·26	− 59·44	− 59·62

TABLE II.

CENTIGRADE SCALE TO FAHRENHEIT
Near the Boiling Point.

Centigrade.	·0	·1	·2	·3	·4	·5	·6	·7	·8	·9
°	F. °	F. °	F. °	F. °	F. °	F. °	F. °	F. °	F. °	F. °
100	212·00	212·18	212·36	212·54	212·72	212·90	213·08	213·26	213·44	213·62
99	210·20	210·38	210·56	210·74	210·92	211·10	211·28	211·46	211·64	211·82
98	208·40	208·58	208·76	208·94	209·12	209·30	209·48	209·66	209·84	210·02
97	206·60	206·78	206·96	207·14	207·32	207·50	207·68	207·86	208·04	208·22
96	204·80	204·98	205·16	205·34	205·52	205·70	205·88	206·06	206·24	206·42
95	203·00	203·18	203·36	203·54	203·72	203·90	204·08	204·26	204·44	204·62
94	201·20	201·38	201·56	201·74	201·92	202·10	202·28	202·46	202·64	202·82
93	199·40	199·58	199·76	199·94	200·12	200·30	200·48	200·66	200·84	201·02
92	197·60	197·78	197·96	198·14	198·32	198·50	198·68	198·86	199·04	199·22
91	195·80	195·98	196·16	196·34	196·52	196·70	196·88	197·06	197·24	197·42
90	194·00	194·18	194·36	194·54	194·72	194·90	195·08	195·26	195·44	195·62

TABLE III.

DIFFERENCES CENTIGRADE TO DIFFERENCES FAHRENHEIT.

Centigrade.	·0	·1	·2	·3	·4	·5	·6	·7	·8	·9
°	F. °	F. °	F. °	F. °	F. °	F. °	F. °	F. °	F. °	F. °
0	0·00	0·18	0·36	0·54	0·72	0·90	1·08	1·26	1·44	1·62
1	1·80	1·98	2·16	2·34	2·52	2·70	2·88	3·06	3·24	3·42
2	3·60	3·78	3·96	4·14	4·32	4·50	4·68	4·86	5·04	5·22
3	5·40	5·58	5·76	5·94	6·12	6·30	6·48	6·66	6·84	7·02
4	7·20	7·38	7·56	7·74	7·92	8·10	8·28	8·46	8·64	8·82
5	9·00	9·18	9·36	9·54	9·72	9·90	10·08	10·26	10·44	10·62
6	10·80	10·98	11·16	11·34	11·52	11·70	11·88	12·06	12·24	12·42
7	12·60	12·78	12·96	13·14	13·32	13·50	13·68	13·86	14·04	14·22
8	14·40	14·58	14·76	14·94	15·12	15·30	15·48	15·66	15·84	16·02
9	16·20	16·38	16·56	16·74	16·92	17·10	17·28	17·46	17·64	17·82

TABLE IV.

MILLIMETRES INTO INCHES.

1 mm. = 0·03937 inch.

Millimetres.	0	1	2	3	4	5	6	7	8	9
	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.
0	0·0000	0·0394	0·0787	0·1181	0·1575	0·1968	0·2362	0·2756	0·3150	0·3543
10	0·3937	0·4331	0·4724	0·5118	0·5512	0·5906	0·6299	0·6693	0·7087	0·7480
20	0·7874	0·8268	0·8661	0·9055	0·9449	0·9842	1·0236	1·0630	1·1024	1·1417
30	1·1811	1·2205	1·2598	1·2992	1·3386	1·3780	1·4173	1·4567	1·4961	1·5354
40	1·5748	1·6142	1·6535	1·6929	1·7323	1·7716	1·8110	1·8504	1·8898	1·9291
50	1·9685	2·0079	2·0472	2·0866	2·1260	2·1654	2·2047	2·2441	2·2835	2·3228
60	2·3622	2·4016	2·4409	2·4803	2·5197	2·5590	2·5984	2·6378	2·6772	2·7165
70	2·7559	2·7953	2·8346	2·8740	2·9134	2·9528	2·9921	3·0315	3·0709	3·1102
80	3·1496	3·1890	3·2283	3·2677	3·3071	3·3464	3·3858	3·4252	3·4646	3·5039
90	3·5433	3·5828	3·6220	3·6614	3·7008	3·7402	3·7795	3·8189	3·8583	3·8976
100	3·9370	3·9764	4·0157	4·0551	4·0945	4·1338	4·1732	4·2126	4·2520	4·2913
110	4·3307	4·3701	4·4094	4·4488	4·4882	4·5276	4·5669	4·6063	4·6457	4·6850
120	4·7244	4·7638	4·8031	4·8425	4·8819	4·9212	4·9606	5·0000	5·0394	5·0787
130	5·1181	5·1575	5·1968	5·2362	5·2756	5·3150	5·3543	5·3937	5·4331	5·4724
140	5·5118	5·5512	5·5905	5·6299	5·6693	5·7086	5·7480	5·7874	5·8268	5·8661
150	5·9055	5·9449	5·9842	6·0236	6·0630	6·1024	6·1417	6·1811	6·2205	6·2598
160	6·2992	6·3386	6·3779	6·4173	6·4567	6·4960	6·5354	6·5748	6·6142	6·6535
170	6·6929	6·7323	6·7716	6·8110	6·8504	6·8898	6·9291	6·9685	7·0079	7·0472
180	7·0866	7·1260	7·1653	7·2047	7·2441	7·2834	7·3228	7·3622	7·4016	7·4409
190	7·4803	7·5197	7·5590	7·5984	7·6378	7·6772	7·7165	7·7559	7·7953	7·8346
200	7·8740	7·9134	7·9527	7·9921	8·0315	8·0708	8·1102	8·1496	8·1890	8·2283
210	8·2677	8·3071	8·3464	8·3858	8·4252	8·4646	8·5039	8·5433	8·5827	8·6220
220	8·6614	8·7008	8·7401	8·7795	8·8189	8·8582	8·8976	8·9370	8·9764	9·0157
230	9·0551	9·0945	9·1338	9·1732	9·2126	9·2520	9·2913	9·3307	9·3701	9·4094
240	9·4488	9·4882	9·5275	9·5669	9·6063	9·6456	9·6850	9·7244	9·7638	9·8031
250	9·8425	9·8819	9·9212	9·9606	10·0000	10·0394	10·0787	10·1181	10·1575	10·1968
260	10·2362	10·2756	10·3149	10·3543	10·3937	10·4330	10·4724	10·5118	10-5512	10·5905
270	10·6299	10·6693	10·7086	10·7480	10·7874	10·8268	10·8661	10·9055	10·9449	10·9842
280	11·0236	11·0630	11·1023	11·1417	11·1811	11·2204	11·2598	11·2992	11·3338	11·3779
290	11·4173	11·4568	11·4960	11·5354	11·5748	11·6142	11·6535	11·6929	11·7323	11·7716
300	11·8110	11·8504	11·8897	11·9291	11·9685	12·0078	12·0472	12·0866	12·1260	12·1653
310	12·2047	12·2441	12·2834	12·3228	12·3622	12·4016	12·4409	12·4803	12·5197	12·5590
320	12·5984	12·6378	12·6771	12·7165	12·7559	12·7952	12·8346	12·8740	12·9134	12·9527
330	12·9921	13·0315	13·0708	13·1102	13·1496	13·1890	13·2283	13·2677	13·3071	13·3464
340	13·3858	13·4252	13·4645	13·5039	13·5433	13·5826	13·6220	13·6614	13·7008	13·7401
350	13·7795	13·8189	13·8582	13·8976	13·9370	13·9764	14·0157	14·0551	14·0945	14·1338
360	14·1732	14·2126	14·2519	14·2913	14·3307	14·3700	14·4094	14·4488	14·4882	14·5275
370	14·5669	14·6063	14·6456	14·6850	14·7244	14·7638	14·8031	14·8425	14·8819	14·9212
380	14·9606	15·0000	15·0393	15·0787	15·1181	15·1574	15·1968	15·2362	15·2756	15·3149
390	15·3543	15·3937	15·4330	15·4724	15·5118	15·5512	15·5905	15·6299	15·6693	15·7086
400	15·7480	15·7874	15·8267	15·8661	15·9055	15·9448	15·9842	16·0236	16·0630	16·1023

Tenths of a millimetre.				Hundredths of a millimetre.			
mm.	Inch.	mm.	Inch.	mm.	Inch.	mm.	Inch.
0·1	0·0039	0·6	0·0236	0·01	0·0004	0·06	0·0024
·2	·0079	·7	·0276	·02	·0008	·07	·0028
·3	·0118	·8	·0315	·03	·0012	·08	·0031
·4	·0157	·9	·0354	·04	·0016	·09	·0035
·5	·0197	1·0	·0394	·05	·0020	·10	·0039

TABLE IV.

MILLIMETRES INTO INCHES.

1 mm. = 0·03937 inch.

Millimetres.	0	1	2	3	4	5	6	7	8	9
	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.
400	15·748	15·787	15·827	15·866	15·905	15·945	15·984	16·024	16·063	16·102
410	16·142	16·181	16·220	16·260	16·299	16·339	16·378	16·417	16·457	16·496
420	16·535	16·575	16·614	16·654	16·693	16·732	16·772	16·811	16·850	16·890
430	16·929	16·968	17·008	17·047	17·087	17·126	17·165	17·205	17·244	17·283
440	17·323	17·362	17·402	17·441	17·480	17·520	17·559	17·598	17·638	17·677
450	17·717	17·756	17·795	17·835	17·874	17·913	17·953	17·992	18·031	18·071
460	18·110	18·150	18·189	18·228	18·268	18·307	18·346	18·386	18·425	18·465
470	18·504	18·543	18·583	18·622	18·661	18·701	18·740	18·779	18·819	18·858
480	18·898	18·937	18·976	19·016	19·055	19·094	19·134	19·173	19·213	19·252
490	19·291	19·331	19·370	19·409	19·449	19·488	19·528	19·567	19·606	19·646
500	19·685	19·724	19·764	19·803	19·842	19·882	19·921	19·961	20·000	20·039
510	20·079	20·118	20·157	20·197	20·236	20·276	20·315	20·354	20·394	20·433
520	20·472	20·512	20·551	20·591	20·630	20·669	20·709	20·748	20·787	20·827
530	20·866	20·905	20·945	20·984	21·024	21·063	21·102	21·142	21·181	21·220
540	21·260	21·299	21·399	21·378	21·417	21·457	21·496	21·535	21·575	21·614
550	21·654	21·693	21·732	21·772	21·811	21·850	21·890	21·929	21·968	22·008
560	22·047	22·087	22·126	22·165	22·205	22·244	22·283	22·323	22·362	22·402
570	22·441	22·480	22·520	22·559	22·598	22·638	22·677	22·716	22·756	22·795
580	22·835	22·874	22·913	22·953	22·992	23·031	23·071	23·110	23·150	23·189
590	23·228	23·268	23·307	23·346	23·386	23·425	23·465	23·504	23·543	23·583
600	23·622	23·661	23·701	23·740	23·779	23·819	23·858	23·898	23·937	23·976
610	24·016	24·055	24·094	24·134	24·173	24·213	24·252	24·291	24·331	24·370
620	24·409	24·449	24·488	24·528	24·567	24·606	24·646	24·685	24·724	24·764
630	24·803	24·842	24·882	24·921	24·961	25·000	25·039	25·079	25·118	25·157
640	25·197	25·236	25·276	25·315	25·354	25·394	25·433	25·472	25·512	25·551
650	25·591	25·630	25·669	25·709	25·748	25·787	25·827	25·866	25·905	25·945
660	25·984	26·024	26·063	26·102	26·142	26·181	26·220	26·260	26·299	26·339
670	26·378	26·417	26·457	26·496	26·535	26·575	26·614	26·653	26·693	26·732
680	26·772	26·811	26·850	26·890	26·929	26·968	27·008	27·047	27·087	27·126
690	27·165	27·205	27·244	27·283	27·323	27·362	27·402	27·441	27·480	27·520
700	27·559	27·598	27·638	27·677	27·716	27·756	27·795	27·835	27·874	27·913
710	27·953	27·992	28·031	28·071	28·110	28·150	28·189	28·228	28·268	28·307
720	28·346	28·386	28·425	28·465	28·504	28·543	28·583	28·622	28·661	28·701
730	28·740	28·779	28·819	28·858	28·898	28·937	28·976	29·016	29·055	29·094
740	29·134	29·173	29·213	29·252	29·291	29·331	29·370	29·409	29·449	29·488
750	29·528	29·567	29·606	29·646	29·685	29·724	29·764	29·803	29·842	29·882
760	29·921	29·961	30·000	30·039	30·079	30·118	30·157	30·197	30·236	30·276
770	30·315	30·354	30·394	30·433	30·472	30·512	30·551	30·590	30·630	30·669
780	30·709	30·748	30·787	30·827	30·866	30·905	30·945	30·984	31·024	31·063
790	31·102	31·142	31·181	31·220	31·260	31·299	31·339	31·378	31·417	31·457
800	31·496	31·535	31·575	31·614	31·653	31·693	31·732	31·772	31·811	31·850
810	31·890	31·929	31·968	32·008	32·047	32·087	32·126	32·165	32·205	32·244
820	32·283	32·323	32·362	32·402	32·441	32·480	32·520	32·559	32·598	32·638
830	32·677	32·716	32·756	32·795	32·835	32·874	32·913	32·953	32·992	33·031
840	33·071	33·110	33·150	33·189	33·228	33·268	33·307	33·346	33·386	33·425
850	33·464	33·504	33·543	33·583	33·622	33·661	33·701	33·740	33·779	33·819
860	33·858	33·898	33·937	33·976	34·016	34·055	34·094	34·134	34·173	34·213
870	34·252	34·291	34·331	34·370	34·409	34·449	34·488	34·527	34·567	34·606
880	34·646	34·685	34·724	34·764	34·803	34·842	34·882	34·921	34·961	35·000
890	35·039	35·079	35·118	35·157	35·197	35·236	35·276	35·315	35·354	35·394

TABLE IV.

MILLIMETRES INTO INCHES.

1 mm. = 0·03937 inch.

Millimetres.	0	1	2	3	4	5	6	7	8	9
	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.
900	35·433	35·472	35·512	35·551	35·590	35·630	35·669	35·709	35·748	35·787
910	35·827	35·866	35·905	35·945	35·984	36·024	36·063	36·102	36·142	36·181
920	36·220	36·260	36·299	36·339	36·378	36·417	36·457	36·496	36·535	36·575
930	36·614	36·653	36·693	36·732	36·772	36·811	36·850	36·890	36·929	36·968
940	37·008	37·047	37·087	37·126	37·165	37·204	37·244	37·283	37·323	37·362
950	37·402	37·441	37·480	37·520	37·559	37·598	37·638	37·677	37·716	37·756
960	37·795	37·835	37·874	37·913	37·953	37·992	38·031	38·071	38·110	38·150
970	38·189	38·228	38·268	38·307	38·346	38·386	38·425	38·464	38·504	38·543
980	38·583	38·622	38·661	38·701	38·740	38·780	38·819	38·858	38·898	38·937
990	38·976	39·016	39·055	39·094	39·134	39·173	39·213	39·252	39·291	39·331
1000	39·370									

TABLE V.

METRES INTO FEET.

1 metre = 39·3700 inches = 3·280833 feet.

Metres.	0	1	2	3	4	5	6	7	8	9
	Feet.	Feet.	Feet.	Feet.	Feet.	Feet.	Feet.	Feet.	Feet.	Feet.
0	0·00	3·28	6·56	9·84	13·12	16·40	19·68	22·97	26·25	29·53
10	32·81	36·09	39·37	42·65	45·93	49·21	52·49	55·77	59·05	62·34
20	65·62	68·90	72·18	75·46	78·74	82·02	85·30	88·58	91·86	95·14
30	98·42	101·71	104·99	108·27	111·55	114·83	118·11	121·39	124·67	127·95
40	131·23	134·51	137·79	141·08	144·36	147·64	150·92	154·20	157·48	160·76
50	164·04	167·32	170·60	173·88	177·16	180·45	183·73	187·01	190·29	193·57
60	196·85	200·13	203·41	206·69	209·97	213·25	216·53	219·82	223·10	226·38
70	229·66	232·94	236·22	239·50	242·78	246·06	249·34	252·62	255·90	259·19
80	262·47	265·75	269·03	272·31	275·59	278·87	282·15	285·43	288·71	291·99
90	295·27	298·56	301·84	305·12	308·40	311·68	314·96	318·24	321·52	324·80
100	328·08	331·36	334·64	337·93	341·21	344·49	347·77	351·05	354·33	357·61
110	360·89	364·17	367·45	370·73	374·01	377·30	380·58	383·86	387·14	390·42
120	393·70	396·98	400·26	403·54	406·82	410·10	413·38	416·67	419·95	423·23
130	426·51	429·79	433·07	436·35	439·63	442·91	446·19	449·47	452·75	456·04
140	459·32	462·60	465·88	469·16	472·44	475·72	479·00	482·28	485·56	488·84
150	492·12	495·41	498·69	501·97	505·25	508·53	511·81	515·09	518·37	521·65
160	524·93	528·21	531·49	534·78	538·06	541·34	544·62	547·90	551·18	554·46
170	557·74	561·02	564·30	567·58	570·86	574·15	577·43	580·71	583·99	587·27
180	590·55	593·83	597·11	600·39	603·67	606·95	610·23	613·52	616·80	620·08
190	623·36	626·64	629·92	633·20	636·48	639·76	643·04	646·32	649·60	652·89
200	656·17	659·45	662·73	666·01	669·29	672·57	675·85	679·13	682·41	685·69
210	688·97	692·26	695·54	698·82	702·10	705·38	708·66	711·94	715·22	718·50
220	721·78	725·06	728·34	731·63	734·91	738·19	741·47	744·75	748·03	751·31
230	754·59	757·87	761·15	764·43	767·71	771·00	774·28	777·56	780·84	784·12
240	787·40	790·68	793·96	797·24	800·52	803·80	807·08	810·37	813·65	816·93
250	820·21	823·49	826·77	830·05	833·33	836·61	839·89	843·17	846·45	849·74
260	853·02	856·30	859·58	862·86	866·14	869·42	872·70	875·98	879·26	882·54
270	885·82	889·11	892·39	895·67	898·95	902·23	905·51	908·79	912·07	915·35
280	918·63	921·91	925·19	928·48	931·76	935·04	938·32	941·60	944·88	948·16
290	951·44	954·72	958·00	961·28	964·56	967·85	971·13	974·41	977·69	980·97
300	984·25	987·53	990·81	994·09	997·37	1000·65	1003·93	1007·22	1010·50	1013·78
310	1017·06	1020·34	1023·62	1026·90	1030·18	1033·46	1036·74	1040·02	1043·30	1046·59
320	1049·87	1053·15	1056·43	1059·71	1062·99	1066·27	1069·55	1072·83	1076·11	1079·39
330	1082·67	1085·96	1089·24	1092·52	1095·80	1099·08	1102·36	1105·64	1109·92	1112·20
340	1115·48	1118·76	1122·04	1125·33	1128·61	1131·89	1135·17	1138·45	1141·73	1145·01
350	1148·29	1151·57	1154·85	1158·13	1161·41	1164·70	1167·98	1171·26	1174·54	1177·82
360	1181·10	1184·38	1187·66	1190·94	1194·22	1197·50	1200·78	1204·07	1207·35	1210·63
370	1213·91	1217·19	1220·47	1223·75	1227·03	1230·31	1233·59	1236·87	1240·15	1243·44
380	1246·72	1250·00	1253·28	1256·56	1259·84	1263·12	1266·40	1269·68	1272·96	1276·24
390	1279·52	1282·81	1286·09	1289·37	1292·65	1295·93	1299·21	1302·49	1305·77	1309·05
400	1312·33	1315·61	1318·89	1322·18	1325·46	1328·74	1332·02	1335·30	1338·58	1341·86
410	1345·14	1348·42	1351·70	1354·98	1358·26	1361·55	1364·83	1368·11	1371·39	1374·67
420	1377·95	1381·23	1384·51	1387·79	1391·07	1394·35	1397·63	1400·92	1404·20	1407·48
430	1410·76	1414·04	1417·32	1420·60	1423·88	1427·16	1430·44	1433·72	1437·00	1440·29
440	1443·57	1446·85	1450·13	1453·41	1456·69	1459·97	1463·25	1466·53	1469·81	1473·09
450	1476·37	1479·66	1482·94	1486·22	1489·50	1492·78	1496·06	1499·34	1502·62	1505·90
460	1509·18	1512·46	1515·74	1519·03	1522·31	1525·59	1528·87	1532·15	1535·43	1538·71
470	1541·99	1545·27	1548·55	1551·83	1555·11	1558·40	1561·68	1564·96	1568·24	1571·52
480	1574·80	1578·08	1581·36	1584·64	1587·92	1591·20	1594·48	1597·77	1601·05	1604·33
490	1607·61	1610·89	1614·17	1617·45	1620·73	1624·01	1627·29	1630·57	1633·85	1637·14
500	1640·42	1643·70	1646·98	1650·26	1653·54	1656·82	1660·10	1663·38	1666·66	1669·94

TABLE V.

METRES INTO FEET.

1 metre = 39·3700 inches = 3·280833 feet.

Metres.	0	10	20	30	40	50	60	70	80	90
	Feet.	Feet.	Feet.	Feet.	Feet.	Feet.	Feet.	Feet.	Feet.	Feet.
500	1640·4	1673·2	1706·0	1738·8	1771·6	1804·5	1837·3	1870·1	1902·9	1935·7
600	1968·5	2001·3	2034·1	2066·9	2099·7	2132·5	2165·3	2198·2	2231·0	2263·8
700	2296·6	2329·4	2362·2	2395·0	2427·8	2460·6	2493·4	2526·2	2559·0	2591·9
800	2624·7	2657·5	2690·3	2723·1	2755·9	2788·7	2821·5	2854·3	2887·1	2919·9
900	2952·7	2985·6	3018·4	3051·2	3084·0	3116·8	3149·6	3182·4	3215·2	3248·0
1000	3280·8	3313·6	3346·4	3379·3	3412·1	3444·9	3477·7	3510·5	3543·3	3576·1
1100	3608·9	3641·7	3674·5	3707·3	3740·1	3773·0	3805·8	3838·6	3871·4	3904·2
1200	3937·0	3969·8	4002·6	4035·4	4068·2	4101·0	4133·8	4166·7	4199·5	4232·3
1300	4265·1	4297·9	4330·7	4363·5	4396·3	4429·1	4461·9	4494·7	4527·5	4560·4
1400	4593·2	4626·0	4658·8	4691·6	4724·4	4757·2	4790·0	4822·8	4855·6	4888·4
1500	4921·2	4954·1	4986·9	5019·7	5052·5	5085·3	5118·1	5150·9	5183·7	5216·5
1600	5249·3	5282·1	5314·9	5347·8	5380·6	5413·4	5446·2	5479·0	5511·8	5544·6
1700	5577·4	5610·2	5643·0	5675·8	5708·6	5741·5	5774·3	5807·1	5839·9	5872·7
1800	5905·5	5938·3	5971·1	6003·9	6036·7	6069·5	6102·3	6135·2	6168·0	6200·8
1900	6233·6	6266·4	6299·2	6332·0	6364·8	6397·6	6430·4	6463·2	6496·0	6528·9
2000	6561·7	6594·5	6627·3	6660·1	6692·9	6725·7	6758·5	6791·3	6824·1	6856·9
2100	6889·7	6922·6	6955·4	6988·2	7021·0	7053·8	7086·6	7119·4	7152·2	7185·0
2200	7217·8	7250·6	7283·4	7316·3	7349·1	7381·9	7414·7	7447·5	7480·3	7513·1
2300	7545·9	7578·7	7611·5	7644·3	7677·1	7710·0	7742·8	7775·6	7808·4	7841·2
2400	7874·0	7906·8	7939·6	7972·4	8005·2	8038·0	8070·8	8103·7	8136·5	8169·3
2500	8202·1	8234·9	8267·7	8300·5	8333·3	8366·1	8398·9	8431·7	8464·5	8497·4
2600	8530·2	8563·0	8595·8	8628·6	8661·4	8694·2	8727·0	8759·8	8792·6	8825·4
2700	8858·2	8891·1	8923·9	8956·7	8989·5	9022·3	9055·1	9087·9	9120·7	9153·5
2800	9186·3	9219·1	9251·9	9284·8	9317·6	9350·4	9383·2	9416·0	9448·8	9481·6
2900	9514·4	9547·2	9580·0	9612·8	9645·6	9678·5	9711·3	9744·1	9776·9	9809·7
3000	9842·5	9875·3	9908·1	9940·9	9973·7	10006·5	10039·3	10072·2	10105·0	10137·8
3100	10170·6	10203·4	10236·2	10269·0	10301·8	10334·6	10367·4	10400·2	10433·0	10465·9
3200	10498·7	10531·5	10564·3	10597·1	10629·9	10662·7	10695·5	10728·3	10761·1	10793·9
3300	10826·7	10859·6	10892·4	10925·2	10958·0	10990·8	11023·6	11056·4	11089·2	11122·0
3400	11154·8	11187·6	11220·4	11253·3	11286·1	11318·9	11351·7	11384·5	11417·3	11450·1
3500	11482·9	11515·7	11548·5	11581·3	11614·1	11647·0	11679·8	11712·6	11745·4	11778·2
3600	11811·0	11843·8	11876·6	11909·4	11942·2	11975·0	12007·8	12040·7	12073·5	12106·3
3700	12139·1	12171·9	12204·7	12237·5	12270·3	12303·1	12335·9	12368·7	12401·5	12434·4
3800	12467·2	12500·0	12532·8	12565·6	12598·4	12631·2	12664·0	12696·8	12729·6	12762·4
3900	12795·2	12828·1	12860·9	12893·7	12926·5	12959·3	12992·1	13024·9	13057·7	13090·5
4000	13123·3	13156·1	13188·9	13221·8	13254·6	13287·4	13320·2	13353·0	13385·8	13418·6
4100	13451·4	13484·2	13517·0	13549·8	13582·6	13615·5	13648·3	13681·1	13713·9	13746·7
4200	13779·5	13812·3	13845·1	13877·9	13910·7	13943·5	13976·3	14009·2	14042·0	14074·8
4300	14107·6	14140·4	14173·2	14206·0	14238·8	14271·6	14304·4	14337·2	14370·0	14402·9
4400	14435·7	14468·5	14501·3	14534·1	14566·9	14599·7	14632·5	14665·3	14698·1	14730·9
4500	14763·7	14796·6	14829·4	14862·2	14895·0	14927·8	14960·6	14993·4	15026·2	15059·0
4600	15091·8	15124·6	15157·4	15190·3	15223·1	15255·9	15288·7	15321·5	15354·3	15387·1
4700	15419·9	15452·7	15485·5	15518·3	15551·1	15584·0	15616·8	15649·6	15682·4	15715·2
4800	15748·0	15780·8	15813·6	15846·4	15879·2	15912·0	15944·8	15977·7	16010·5	16043·3
4900	16076·1	16108·9	16141·7	16174·5	16207·3	16240·1	16272·9	16305·7	16338·5	16371·4
5000	16404·2	16437·0	16469·8	16502·6	16535·4	16568·2	16601·0	16633·8	16666·6	16699·4

Tenths of a metre,	0·1	0·2	0·3	0·4	0·5	0·6	0·7	0·8	0·9
Feet, - - -	0·328	0·656	0·984	1·312	1·640	1·968	2·297	2·625	2·953

TABLE VI.

KILOMETRES INTO MILES.

1 kilometre = 0·621370 mile.

Kilometres.	0	1	2	3	4	5	6	7	8	9
	Miles.	Miles.	Miles.	Miles.	Miles.	Miles.	Miles.	Miles.	Miles.	Miles.
0	0·0	0·6	1·2	1·9	2·5	3·1	3·7	4·3	5·0	5·6
10	6·2	6·8	7·5	8·1	8·7	9·3	9·9	10·6	11·2	11·8
20	12·4	13·0	13·7	14·3	14·9	15·5	16·2	16·8	17·4	18·0
30	18·6	19·3	19·9	20·5	21·1	21·7	22·4	23·0	23·6	24·2
40	24·9	25·5	26·1	26·7	27·3	28·0	28·6	29·2	29·8	30·4
50	31·1	31·7	32·3	32·9	33·6	34·2	34·8	35·4	36·0	36·7
60	37·3	37·9	38·5	39·1	39·8	40·4	41·0	41·6	42·3	42·9
70	43·5	44·1	44·7	45·4	46·0	46·6	47·2	47·8	48·5	49·1
80	49·7	50·3	51·0	51·6	52·2	52·8	53·4	54·1	54·7	55·3
90	55·9	56·5	57·2	57·8	58·4	59·0	59·7	60·3	60·9	61·5
100	62·1	62·8	63·4	64·0	64·6	65·2	65·9	66·5	67·1	67·7
110	68·4	69·0	69·6	70·2	70·8	71·5	72·1	72·7	73·3	73·9
120	74·6	75·2	75·8	76·4	77·0	77·7	78·3	78·9	79·5	80·2
130	80·8	81·4	82·0	82·6	83·3	83·9	84·5	85·1	85·7	86·4
140	87·0	87·6	88·2	88·9	89·5	90·1	90·7	91·3	92·0	92·6
150	93·2	93·8	94·4	95·1	95·7	96·3	96·9	97·6	98·2	98·8
160	99·4	100·0	100·7	101·3	101·9	102·5	103·1	103·8	104·4	105·0
170	105·6	106·3	106·9	107·5	108·1	108·7	109·4	110·0	110·6	111·2
180	111·8	112·5	113·1	113·7	114·3	115·0	115·6	116·2	116·8	117·4
190	118·1	118·7	119·3	119·9	120·5	121·2	121·8	122·4	123·0	123·7
200	124·3	124·9	125·5	126·1	126·8	127·4	128·0	128·6	129·2	129·9
210	130·5	131·1	131·7	132·4	133·0	133·6	134·2	134·8	135·5	136·1
220	136·7	137·3	137·9	138·6	139·2	139·8	140·4	141·1	141·7	142·3
230	142·9	143·5	144·2	144·8	145·4	146·0	146·6	147·3	147·9	148·5
240	149·1	149·8	150·4	151·0	151·6	152·2	152·9	153·5	154·1	154·7
250	155·3	156·0	156·6	157·2	157·8	158·4	159·1	159·7	160·3	160·9
260	161·6	162·2	162·8	163·4	164·0	164·7	165·3	165·9	166·5	167·1
270	167·8	168·4	169·0	169·6	170·3	170·9	171·5	172·1	172·7	173·4
280	174·0	174·6	175·2	175·8	176·5	177·1	177·7	178·3	179·0	179·6
290	180·2	180·8	181·4	182·1	182·7	183·3	183·9	184·5	185·2	185·8
300	186·4	187·0	187·7	188·3	188·9	189·5	190·1	190·8	191·4	192·0
310	192·6	193·2	193·9	194·5	195·1	195·7	196·4	197·0	197·6	198·2
320	198·8	199·5	200·1	200·7	201·3	201·9	202·6	203·2	203·8	204·4
330	205·1	205·7	206·3	206·9	207·5	208·2	208·8	209·4	210·0	210·6
340	211·3	211·9	212·5	213·1	213·8	214·4	215·0	215·6	216·2	216·9
350	217·5	218·1	218·7	219·3	220·0	220·6	221·2	221·8	222·5	223·1
360	223·7	224·3	224·9	225·6	226·2	226·8	227·4	228·0	228·7	229·3
370	229·9	230·5	231·1	231·8	232·4	233·0	233·6	234·3	234·9	235·5
380	236·1	236·7	237·4	238·0	238·6	239·2	239·8	240·5	241·1	241·7
390	242·3	243·0	243·6	244·2	244·8	245·4	246·1	246·7	247·3	247·9
400	248·5	249·2	249·8	250·4	251·0	251·7	252·3	252·9	253·5	254·1
410	254·8	255·4	256·0	256·6	257·2	257·9	258·5	259·1	259·7	260·4
420	261·0	261·6	262·2	262·8	263·5	264·1	264·7	265·3	265·9	266·6
430	267·2	267·8	268·4	269·1	269·7	270·3	270·9	271·5	272·2	272·8
440	273·4	274·0	274·6	275·3	275·9	276·5	277·1	277·8	278·4	279·0

TABLE VI.

KILOMETRES INTO MILES.

1 kilometre = 0·621370 mile.

Kilo-metres.	0	1	2	3	4	5	6	7	8	9
	Miles.	Miles.	Miles.	Miles.	Miles.	Miles.	Miles.	Miles.	Miles.	Miles.
450	279·6	280·2	280·9	281·5	282·1	282·7	283·3	284·0	284·6	285·2
460	285·8	286·5	287·1	287·7	288·3	288·9	289·6	290·2	290·8	291·4
470	292·0	292·7	293·3	293·9	294·5	295·2	295·8	296·4	297·0	297·6
480	298·3	298·9	299·5	300·1	300·7	301·4	302·0	302·6	303·2	303·8
490	304·5	305·1	305·7	306·3	307·0	307·6	308·2	308·8	309·4	310·1
500	310·7	311·3	311·9	312·5	313·2	313·8	314·4	315·0	315·7	316·3
510	316·9	317·5	318·1	318·8	319·4	320·0	320·6	321·2	321·9	322·5
520	323·1	323·7	324·4	325·0	325·6	326·2	326·8	327·5	328·1	328·7
530	329·3	329·9	330·6	331·2	331·8	332·4	333·1	333·7	334·3	334·9
540	335·5	336·2	336·8	337·4	338·0	338·6	339·3	339·9	340·5	341·1
550	341·8	342·4	343·0	343·6	344·2	344·9	345·5	346·1	346·7	347·3
560	348·0	348·6	349·2	349·8	350·5	351·1	351·7	352·3	352·9	353·6
570	354·2	354·8	355·4	356·0	356·7	357·3	357·9	358·5	359·2	359·8
580	360·4	361·0	361·6	362·3	362·9	363·5	364·1	364·7	365·4	366·0
590	366·6	367·2	367·9	368·5	369·1	369·7	370·3	371·0	371·6	372·2
600	372·8	373·4	374·1	374·7	375·3	375·9	376·6	377·2	377·8	378·4
610	379·0	379·7	380·3	380·9	381·5	382·1	382·8	383·4	384·0	384·6
620	385·2	385·9	386·5	387·1	387·7	388·4	389·0	389·6	390·2	390·8
630	391·5	392·1	392·7	393·3	393·9	394·6	395·2	395·8	396·4	397·1
640	397·7	398·3	398·9	399·5	400·2	400·8	401·4	402·0	402·6	403·3
650	403·9	404·5	405·1	405·8	406·4	407·0	407·6	408·2	408·9	409·5
660	410·1	410·7	411·3	412·0	412·6	413·2	413·8	414·5	415·1	415·7
670	416·3	416·9	417·6	418·2	418·8	419·4	420·0	420·7	421·3	421·9
680	422·5	423·2	423·8	424·4	425·0	425·6	426·3	426·9	427·5	428·1
690	428·7	429·4	430·0	430·6	431·2	431·9	432·5	433·1	433·7	434·3
700	435·0	435·6	436·2	436·8	437·4	438·1	438·7	439·3	439·9	440·6
710	441·2	441·8	442·4	443·0	443·7	444·3	444·9	445·5	446·1	446·8
720	447·4	448·0	448·6	449·3	449·9	450·5	451·1	451·7	452·4	453·0
730	453·6	454·2	454·8	455·5	456·1	456·7	457·3	457·9	458·6	459·2
740	459·8	460·4	461·1	461·7	462·3	462·9	463·5	464·2	464·8	465·4
750	466·0	466·6	467·3	467·9	468·5	469·1	469·8	470·4	471·0	471·6
760	472·2	472·9	473·5	474·1	474·7	475·3	476·0	476·6	477·2	477·8
770	478·5	479·1	479·7	480·3	480·9	481·6	482·2	482·8	483·4	484·0
780	484·7	485·3	485·9	486·5	487·2	487·8	488·4	489·0	489·6	490·3
790	490·9	491·5	492·1	492·7	493·4	494·0	494·6	495·2	495·9	496·5
800	497·1	497·7	498·3	499·0	499·6	500·2	500·8	501·4	502·1	502·7
810	503·3	503·9	504·6	505·2	505·8	506·4	507·0	507·7	508·3	508·9
820	509·5	510·1	510·8	511·4	512·0	512·6	513·3	513·9	514·5	515·1
830	515·7	516·4	517·0	517·6	518·2	518·8	519·5	520·1	520·7	521·3
840	522·0	522·6	523·2	523·8	524·4	525·1	525·7	526·3	526·9	527·5
850	528·2	528·8	529·4	530·0	530·6	531·3	531·9	532·5	533·1	533·8
860	534·4	535·0	535·6	536·2	536·9	537·5	538·1	538·7	539·3	540·0
870	540·6	541·2	541·8	542·5	543·1	543·7	544·3	544·9	545·6	546·2
880	546·8	547·4	548·0	548·7	549·3	549·9	550·5	551·2	551·8	552·4
890	553·0	553·6	554·3	554·9	555·5	556·1	556·7	557·4	558·0	558·6

TABLE VI.

KILOMETRES INTO MILES.

1 kilometre = 0·621370 mile.

Kilo-metres.	0	1	2	3	4	5	6	7	8	9
	Miles.	Miles.	Miles.	Miles.	Miles.	Miles.	Miles.	Miles.	Miles.	Miles.
900	559·2	559·9	560·5	561·1	561·7	562·3	563·0	563·6	564·2	564·8
910	565·4	566·1	566·7	567·3	567·9	568·6	569·2	569·8	570·4	571·0
920	571·7	572·3	572·9	573·5	574·1	574·8	575·4	576·0	576·6	577·3
930	577·9	578·5	579·1	579·7	580·4	581·0	581·6	582·2	582·8	583·5
940	584·1	584·7	585·3	586·0	586·6	587·2	587·8	588·4	589·1	589·7
950	590·3	590·9	591·5	592·2	592·8	593·4	594·0	594·7	595·3	595·9
960	596·5	597·1	597·8	598·4	599·0	599·6	600·2	600·9	601·5	602·1
970	602·7	603·4	604·0	604·6	605·2	605·8	606·5	607·1	607·7	608·3
980	608·9	609·6	610·2	610·8	611·4	612·0	612·7	613·3	613·9	614·5
990	615·2	615·8	616·4	617·0	617·6	618·3	618·9	619·5	620·1	620·7
1000	621·4	622·0	622·6	623·2	623·9	624·5	625·1	625·7	626·3	627·0

km.	Miles.	km.	Miles.	km.	Miles.	km.	Miles.
1000	621·4	6000	3728·2	11000	6835·1	16000	9941·9
2000	1242·7	7000	4349·6	12000	7456·4	17000	10563·3
3000	1864·1	8000	4971·0	13000	8077·8	18000	11184·7
4000	2485·5	9000	5592·3	14000	8699·2	19000	11806·0
5000	3106·8	10000	6213·7	15000	9320·5	20000	12427·4

TABLE VII.

KILOGRAMMES INTO AVOIRDUPOIS POUNDS AND OUNCES

1 kilogramme = 2·204622 avoirdupois pounds.

Kilo-gram's.	0·0	0·1	0·2	0·3	0·4	0·5	0·6	0·7	0·8	0·9
	Av. lbs.	Av. lbs.	Av. lbs.	Av. lbs.	Av. lbs.	Av. lbs.	Av. lbs.	Av. lbs.	Av. lbs.	Av. lbs.
0	0·000	0·220	0·441	0·661	0·882	1·102	1·323	1·543	1·764	1·984
1	2·205	2·425	2·646	2·866	3·086	3·307	3·527	3·748	3·968	4·189
2	4·409	4·630	4·850	5·071	5·291	5·512	5·732	5·952	6·173	6·393
3	6·614	6·834	7·055	7·275	7·496	7·716	7·937	8·157	8·378	8·598
4	8·818	9·039	9·259	9·480	9·700	9·921	10·141	10·362	10·582	10·803
5	11·023	11·244	11·464	11·684	11·905	12·125	12·346	12·566	12·787	13·007
6	13·228	13·448	13·669	13·889	14·110	14·330	14·551	14·771	14·991	15·212
7	15·432	15·653	15·873	16·094	16·314	16·535	16·755	16·976	17·196	17·417
8	17·637	17·857	18·078	18·298	18·519	18·739	18·960	19·180	19·401	19·621
9	19·842	20·062	20·283	20·503	20·723	20·944	21·164	21·385	21·605	21·826

Tenths of a Kilogramme into Ounces.

kg.	Oz.	kg.	Oz.
0·1	3·5274	0·6	21·1644
·2	7·0548	·7	24·6918
·3	10·5822	·8	28·2192
·4	14·1096	·9	31·7466
·5	17·6370	1·0	35·2740

Hundredths of a Kilogramme into Decimals of a Pound and Ounces.

kg.	Av. lbs. Oz.	kg.	Av. lbs. Oz.
0·01	0·022 = 0·35	0·06	0·132 = 2·12
·02	·044 = 0·71	·07	·154 = 2·47
·03	·066 = 1·06	·08	·176 = 2·82
·04	·088 = 1·41	·09	·198 = 3·17
·05	·110 = 1·76	·10	·220 = 3·53

TABLE VIII.

GRAMMES INTO GRAINS.

1 gramme = 15·432351 grains.

Gram's	·0	·1	·2	·3	·4	·5	·6	·7	·8	·9
	Grains.	Grains.	Grains.	Grains.	Grains.	Grains.	Grains	Grains.	Grains.	Grains.
0	0·00	1·54	3·09	4·63	6·17	7·72	9·26	10·80	12·35	13·89
1	15·43	16·98	18·52	20·06	21·61	23·15	24·69	26·24	27·78	29·32
2	30·86	32·41	33·95	35·49	37·04	38·58	40·12	41·67	43·21	44·75
3	46·30	47·84	49·38	50·93	52·47	54·01	55·56	57·10	58·64	60·19
4	61·73	63·27	64·82	66·36	67·90	69·45	70·99	72·53	74·08	75·62
5	77·16	78·71	80·25	81·79	83·33	84·88	86·42	87·96	89·51	91·05
6	92·59	94·14	95·68	97·22	98·77	100·31	101·85	103·40	104·94	106·48
7	108·03	109·57	111·11	112·66	114·20	115·74	117·29	118·83	120·37	121·92
8	123·46	125·00	126·55	128·09	129·63	131·18	132·72	134·26	135·80	137·35
9	138·89	140·43	141·98	143·52	145·06	146·61	148·15	149·69	151·24	152·78

	0	1	2	3	4	5	6	7	8	9
	Grains.	Grains.	Grains.	Grains.	Grains.	Grains.	Grains.	Grains.	Grains.	Grains.
0	0·00	15·43	30·86	46·30	61·73	77·16	92·59	108·03	123·46	138·89
10	154·32	169·76	185·19	200·62	216·05	231·49	246·92	262·35	277·78	293·21
20	308·65	324·08	339·51	354·94	370·38	385·81	401·24	416·67	432·11	447·54
30	462·97	478·40	493·84	509·27	524·70	540·13	555·56	571·00	586·43	601·86
40	617·29	632·73	648·16	663·59	679·02	694·46	709·89	725·32	740·75	756·19
50	771·62	787·05	802·48	817·91	833·35	848·78	864·21	879·64	895·08	910·51
60	925·94	941·37	956·81	972·24	987·67	1003·10	1018·54	1033·97	1049·40	1064·83
70	1080·26	1095·70	1111·13	1126·56	1141·99	1157·43	1172·86	1188·29	1203·72	1219·16
80	1234·59	1250·02	1265·45	1280·89	1296·32	1311·75	1327·18	1342·62	1358·05	1373·48
90	1388·91	1404·34	1419·78	1435·21	1450·64	1466·07	1481·51	1496·94	1512·37	1527·80

gramme.	Grain.	gramme.	Grain.	gramme.	Grain.	gramme.	Grain.
0·01	0·154	0·06	0·926	0·001	0·015	0·006	0·093
·02	·309	·07	1·080	·002	031	·007	·108
·03	·463	·08	1·235	·003	·046	·008	·123
·04	·617	·09	1·389	·004	·062	·009	·139
·05	·772	·10	1·543	·005	·077	·010	·154

TABLE IX.

METRES PER SECOND INTO MILES PER HOUR.

1 metre per second = 2·236932 miles per hour.

Metres per Second	0·0	0·1	0·2	0·3	0·4	0·5	0·6	0·7	0·8	0·9
	Miles per hr.	Miles per hr.	Miles per hr.	Miles per hr.	Miles per hr.	Miles per hr.	Miles per hr.	Miles per hr.	Miles per hr.	Miles per hr.
0	0·0	0·2	0·4	0·7	0·9	1·1	1·3	1·6	1·8	2·0
1	2·2	2·5	2·7	2·9	3·1	3·4	3·6	3·8	4·0	4·3
2	4·5	4·7	4·9	5·1	5·4	5·6	5·8	6·0	6·3	6·5
3	6·7	6·9	7·2	7·4	7·6	7·8	8·1	8·3	8·5	8·7
4	8·9	9·2	9·4	9·6	9·8	10·1	10·3	10·5	10·5	11·0
5	11·2	11·4	11·6	11·9	12·1	12·3	12·5	12·8	13·0	13·2
6	13·4	13·6	13·9	14·1	14·3	14·5	14·8	15·0	15·2	15·4
7	15·7	15·9	16·1	16·3	16·6	16·8	17·0	17·2	17·4	17·7
8	17·9	18·1	18·3	18·6	18·8	19·0	19·2	19·5	19·7	19·9
9	20·1	20·4	20·6	20·8	21·0	21·3	21·5	21·7	21·9	22·1

B. ADDENDA.

The following additional references, mostly of recent date, may be found useful:

Page 29. **City temperatures.**—W. H. Hammon and F. W. Duenckel: "Abstract of a Comparison of the Minimum Temperatures recorded at the U.S. Weather Bureau and the Forest Park Meteorological Observatories, St. Louis, Mo., for the Year 1891," *Mo. Weather Rev.*, XXX., 1902, 12-13.

Page 78. **London fogs.**—R. H. Curtis: "Hourly Variations of Sunshine at Seven Stations in the British Isles," *Quart. Journ. Roy. Met. Soc.*, XXI., 1895, 216-227.

Page 79. **Micro-organisms.**—*Nature*, LXV., 1902, 573.

Pages 100, 200, 217. **Tables.**—W. von Bezold: "Ueber klimatologische Mittelwerthe für ganze Breitenkreise," *Sitzungsber. Berlin. Akad.*, LIII., December, 1901, 1330-1344. Reviewed by J. Hann, *M.Z.*, XIX., 1902, 260-269.

Page 202. **Foot-note.**—J. Hann: "Nochmals die Temperatur der höheren Breiten der südlichen Halbkugel," *M.Z.*, XIII., 1896, 180-183.

Page 238. **Importance of exposure in controlling surface temperature in mountain climates.**—Maurice Lugeon: "Quelques Mots sur le Groupement de la Population du Valais," *Etrennes helvétiques pour* 1902, Lausanne, 1902. (Note in *Science*, *N.S.*, XV., 1902, 915.)

Page 255. **Relation of inversions of temperature to vegetation.**—S. Alexander: "The Thermal Belts and Cold Island of Southeastern Michigan," *Am. Met. Journ.*, I., 1884-85, 467-471. Regarding this article, Mr. Alexander writes under date of March 21, 1902, as follows: "The article states the facts correctly. The thermal belts in question are one of the best peach-growing regions in the country. One grower during the past year shipped 30,000 bushels; another 5000, and so on down to a few hundred bushels each by many other growers. That article was written sixteen years ago, and after observing the belts and island for that length of time I see no reason for changing or erasing a single sentence."

Page 314. **Table, Equatorial limits of snowfall.**—The latitudes for North America as given by Fischer have been changed somewhat, after consultation with Professor A. J. Henry.

Page 318. **The height of the snow-line in different mountains.**—J. Jegerlehner: "Die Schneegrenze in den Gletschergebieten der Schweiz," *Gerland's Beiträge zur Geophysik*, V., 1902, No. 3, 486-566.

Page 334. **Foot-note 2.**—In the discussion entitled "Weitere Untersuchungen ueber die tägliche Oscillation des Barometers," *D. W.A.*, LIX., 1892, 333-337, the author of the present book has tried to answer the objections to his theory which have been urged by Sprung. Moreover, the valley wind of the upper Engadine is itself the best answer to these objections. See R. Billwiller: "Der Thalwind des Oberengadin," *Ann. d. schweiz. Met. Centr.-Anstalt.*, 1893. Summarised in *M.Z.*, XIII., 1896, 129-138.

Page 363. **The bora and the mistral.**—E. Mazelle: "Einfluss der Bora auf die tägliche Periode einiger meteorologischer Elemente," *D. W.A.*, LXXIII., 1901, 67-100.

Page 398. **Arrhenius' theory.**—N. Ekholm: "Ueber Emission und Absorption der Wärme und deren Bedeutung für die Temperatur der Erdoberfläche," *M.Z.*, XIX., 1902, 1-26, 489-505 (to be continued).

Page 404. **Sunspots and meteorological cycles.**—Sir N. and W. J. S. Lockyer: "On some Phenomena which suggest a Short Period of Solar and Meteorological Changes," *Proc. Royal Soc.*, LXX., 1902, 500-504.

Page 405. **Sunspots and temperature.**—Maxwell Hall: "Temperatures in Kingston, Jamaica, and the Connection between the Sunspot Frequency, the Mean Maximum Temperature, and the Rainfall in Jamaica," *Jamaica Weather Review*, No. 275, 1902.

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